SEDIMENTOLOGY AND TECTONICS OF THE COLLISION COMPLEX IN THE EAST ARM OF SULAWESI, INDONESIA

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".... The history of knowledge has been characterised by periodic formulation of hypotheses that generalised most factual information available at a given time. Science is a process of continuous refinement and testing of such generalisations. Hypotheses inevitably have been coloured by the temperaments, expareiences, and prejudices of their advocates, which make them all the more interesting to study"

(Dott and Batten : Evolution of the Earth).

To my lovely girls:

Riri

Vera

Intan

Clara

and my wife,

Nelda.

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ABSTRACT

An imbricated Mesozoic to Palaeogene continental margin sequence is juxtaposed with ophiolitic rocks in the East Arm of Sulawesi, Indonesia. The two tectonic terranes are bounded by the Batui Thrust and Balantak Fault System, which are considered to be the surface expression of the collision zone between the Banggai-Sula Platform and the Eastern Sulawesi Ophiolite Belt. The collision complex contains three distinctive sedimentary sequences : 1) Triassic-Palaeogene continental margin sediments, ii) Cretaceous pelagic sediments and iii) Neogene coarse clastic sediments and volcanogenic turbidites.

(i) Late Triassic Lemo Beds consisting largely of carbonate-slope deposits and subsidiary clastics including quartz-rich lithic sandstones and lensoidal pebbly mudstone and conglomeratic breccia. The hemipelagic limestones are rich in micro-fossils. Some beds of the limestone contain bivalves and ammonites, including Misolia, which typifies the Triassic-Jurassic sequence of eastern Indonesia. The Jurassic Kapali Beds are dominated by quartzose arenites containing significant amounts of plant remains and lumps of coal. The Late Jurassic sediments consist of neritic carbonate deposits (Nambo Beds and Sinsidik Beds) containing ammonites and belemnites, including Belemnopsis uhligi Stevens, of Late Jurassic age. The Jurassic sediments are overlain unconformably by Late Cretaceous Luok Beds which are

predominantly calcilutite with chert nodules rich in microfossils. The Luok Beds are unconformably overlain by the Palaeogene Salodik Limestones which consist of carbonate platform sediments rich in both benthic and planktonic foraminifera of Eocene to Early Miocene age. These sediments were deposited on the continental margin of the Banggai-Sula Platform.

(ii) Deep-sea sediments (Boba Beds) consist largely of chert and subsidiary calcilutite rich in radiolaria of Cretaceous age. These rocks are part of an ophiolite suite.

(iii) Coarse clastic sediments (Kolo Beds and Biak Conglomerates) are typical post-orogenic clastic rocks deposited on top of the collision complex. They are composed of material derived from both the continental margin sequence and ophiolite suite. Volcanogenic Lonsuit Turbidites occur in the northern part of the East Arm in Poh Head and unconformably overlie the ophiolite suite. Late Miocene to Pliocene planktonic foraminifera occur in the intercalated marlstone and marly sandstone beds within these rocks.

The collision zone is marked by the occurrence of Kolokolo Melange, which contain exotic fragments detached from both the ophiolite suite and the continental margin sequence and a matrix of calcareous mudstone and marlstone rich in planktonic foraminifera of late Middle Miocene to Pliocene age. The melange is believed to have been formed during and after the collision of the Banggai-Sula Platform with the Eastern Sulawesi Ophiolite Belt. Hence, the collision event took place in Middle Miocene time. The occurrence of at least three terraces of Quaternary coraline reefs on the south coast of the East Arm of Sulawesi testifies to the rapid uplift of the region. Seismic data suggest that the collision might still be in progress at the present time.

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CHAPTER 1

GENERAL INTRODUCTION

This thesis contains analyses of the sedimentology, petrology and stratigraphy of Triassic, Jurassic, Cretaceous and Tertiary rocks in the East Arm of Sulawesi, Indonesia. It contains also, the result of a study of structural relationships of these sedimentary successions and tectonic setting of the Balantak Ophiolites which form the northern portion of the Eastern Sulawesi Ophiolite Belt (ESOB).

An effort has been made to date the mafic part of the Balantak Ophiolites and the pelagic sediments which covered the ophiolite suite.

The region has been mapped recently by the Indonesian Geological Research and Development Centre (GRDC) ; results have been set forth in the geological reports of Poso, Batui, and Luwuk Quadrangles of 1:250.000 scale. The author was assigned as a senior geologist to this mapping project. Data collected during the project are included in this thesis with the addition of data collected from six selected sections for more detailed study.

The final analysis is based on six avenues of investigations, namely, 1. stratigraphy and biostratigraphy, 2. sedimentology, 3. petrology, 4. basin analysis, 5. structural analysis and 6. tectonic analysis of the region.

1.1. GEOGRAPHIC SETTING

1.1.1. Regional Setting

Sulawesi (Celebes) is the largest island in the central part of the Indonesian archipelago (Fig. 1.1). It

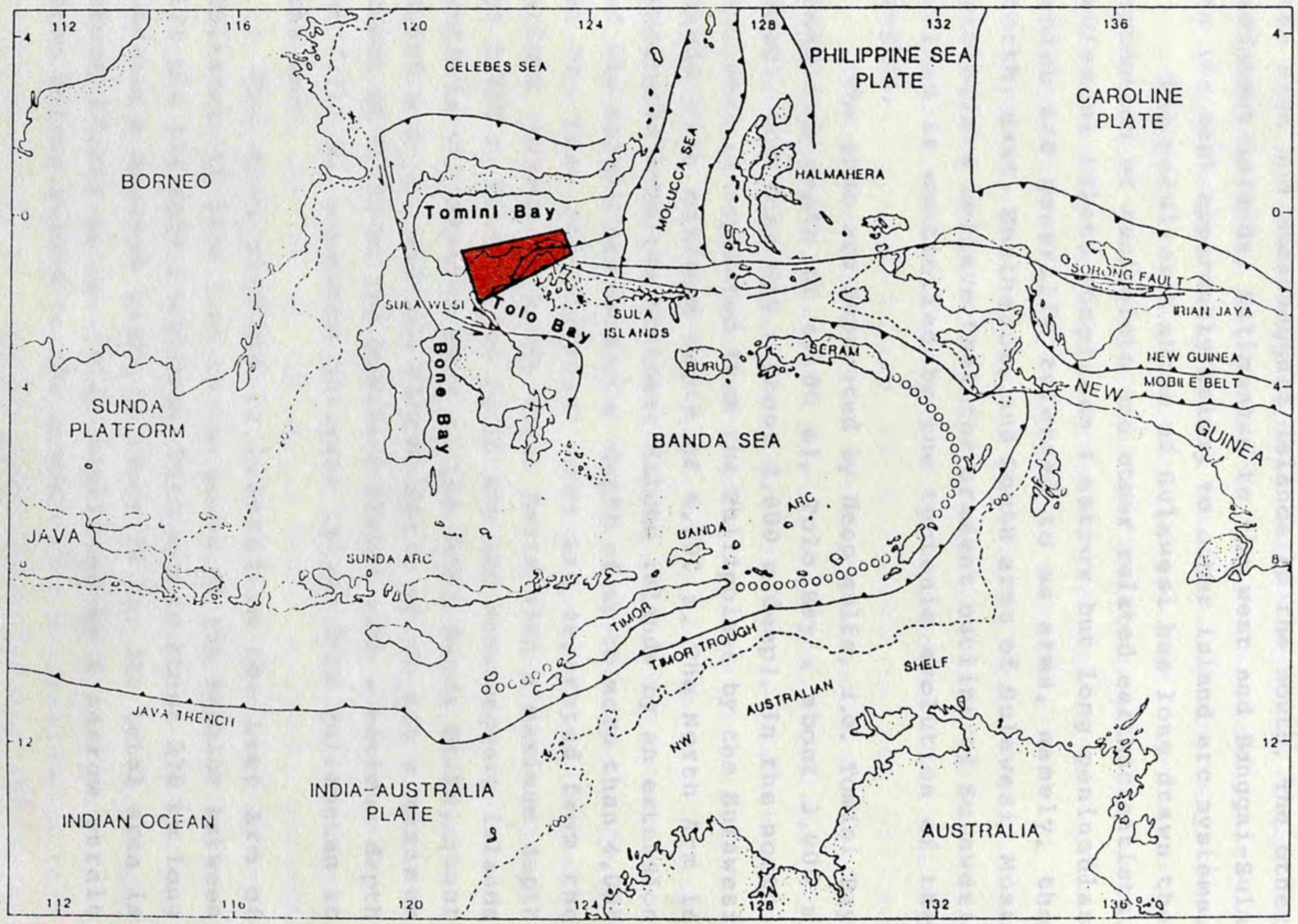


Fig. 1.1 Map showing tectonic configuration of eastern Indonesia (After Hamilton, 1979; Silver, 1979) and the area studied. Teeth along thrusts are on the overriding plate.

is a connecting link between the island arc systems of Sangir-Talaud and Philippines to the north, Banda Arc to the east and Nusatenggara islands to the south. The other adjacent islands, Kalimantan to the west and Banggai-Sula to the east apparently belong to other island arc systems.

The peculiar K shape of Sulawesi has long drawn the attention of geologists and other related earthscientists. Sulawesi island comprises 4 narrow but long peninsulas which are generally referred to as arms, namely, the North, East, Southeast, and South arms of Sulawesi. Most geologists believe that the present outline of Sulawesi island is controlled by the tectonic evolution of the region.

The arms are separated by deep gulfs, i.e. Tomini Bay (maximum depth of 4,180 m), Tolo Bay (about 3,000 m deep), and Bone Bay (about 2,000 m deep). In the north, Sulawesi is separated from the Philippines by the Sulawesi Basin with maximum depth of 6,200 m. The North Arm is separated from the northern Maluku islands by an extension of the Maluku Sea having a depth of up to more than 4,000 m. The East and Southeast Arms are separated from the Maluku islands by North Banda Basin with a maximum depth of 5,750 m. Between the South Arm and Nusatenggara Islands extends the western part of the South Banda Basin, about 4,500 m deep, and the Flores Basin which has a maximum depth of 5,140 m. The Makassar Strait with a maximum depth of 2,717 m, separates Sulawesi Island from Kalimantan to the west.

The area studied is located in the East Arm of Sulawesi. It lies just to the south of the Equator between 121 and 124 East Longitude. This arm is about 270 km long and has a maximum width of about 70 km. Its total area is about 17,250 sq.km. It is separated by a narrow strait from Peleng Island to the south.

1.1.2. Culture

The East Arm of Sulawesi consists of two Kabupaten (Administrative Regencies), the western area is incorporated in the Kabupaten (Regency) of Poso with Poso as its capital and the eastern area and its surroundings islands are under the Kabupaten Luwuk-Banggai with Luwuk as its capital. The whole area is included in the Central Sulawesi Province with Palu as its capital.

The region is sparsely populated, about 15/1 sq.km. The total population of approximately 300,000 is unevenly distributed, and mostly inhabits the coastal areas. At present the East Arm of Sulawesi is one of the transmigration destinations. Existing transmigration sites include Toili, about 60 km west of Luwuk, Lamala, 40 km east of Luwuk and Bunta on the north coast. Migrants come mostly from Jawa and Bali, but there are also local and spontaneous migrants who come from the adjacent areas, such as Gorontalo, Minahasa, Bugis, Badjo and Buton.

The native inhabitants comprise many tribes including the Mori and Pamona in the western part, Saluan and Wana in the middle part, Balantak in the eastern part of the East Arm, and Banggai and Sea-sea in Peleng Island. Each tribe has its own language and customs; most of natives can speak or communicate in Bahasa Indonesia, the national language. The majority of the inhabitants are Moslem but some are Christians. They are farmers and fishermen. Main products are copra and fish. Other products include timber, rottan, and damar. The transmigrant areas produce rice, corn, ubi, cashew nut, vegetables and various fruits. Wildlife includes snakes, deer, crocodiles and other animals found only in Sulawesi Island, such as <u>anoa</u> (a pigmy cow) and the <u>maleo</u> bird.

Like other parts of the Indonesia the weather of this area is tropical, and is influenced by dry and rainy seasons. The dry season lasts from August to January, whereas the rainy season has its peak between March and June. During the rainy months the rivers are in flood and it is not possible to traverse them. Storms prevail at sea, and the huge waves makes sea travel very dangerous.

1.1.3. Accessibility

Luwuk can be reached from either Ujungpandang, Palu or Manado by using Twin Otter or Cessna aircraft of the Merpati Nusantara Airlines (MNA) or by ship operated by the Governments PELNI or by private shipping companies. Luwuk has both air and sea ports. Luwuk is connected by road with Palu via Poso.

Transportation and communication within the area studied is either by sea or by road. Travel by sea uses motorized sea-perahus, boats or vessels. Most of the roads are unsurfaced, and some sections are just recently built. The road is asphalted between Luwuk and Batui, and between Luwuk and Pagimana via Poh. The bridges along the road between Batui and Toili are broken down due to the huge flood in mid-1983. The new road between Toili and Baturube via Kolo Atas is still under construction. A new road from Salodik to Binsil is muddy and slippery in the rainy season. The section between Luwuk and Balantak can be used at any season, but is in a poor state of repair. Many footpaths crossing the East Arm from north to south coast are now hardly used.

1.1.4. Topography

The morphology of the East Arm of Sulawesi can be classified into 4 units, i.e. lowlands, karst, hilly topography and mountain ranges. A low flat plain with an elevation between 0 and 50 m above sea level is found along the coast, from Batui to Kolo Atas on the south coast, along the north coast of the Poh Head, and along the north coast between the Bunta and Balingara and Ampana areas. These are areas composed of river and beach deposits and swamps.

The karst, which is formed from Tertiary limestones (i.e. Salodik Limestones) and Quaternary coralline reefs, occupies the areas north of Batui, the southern coast near Luwuk, the south coast of Poh Head, the western part of Peleng Island and some places on the north coast of the East Arm. The morphology of karst, with an elevation between 50 m and 1000 m above sea level can be observed clearly on aerial photographs and the SLAR imagery. Dolinas and underground streams are commonly found within this unit. Near Luwuk, the karst shows at least 4 terraces with elevations of approximately 20, 75, 200, and 400 metres above sea level.

The morphology of the hilly areas is commonly steep slopes, and undulatory peaks and rather rough relief. Elevation ranges from 50 m to 700 m above sea level. The hills are formed by volcanogenic turbidite sediments and pillow basalt in the northern part of the Poh Head and Togian Islands, by Palaeogene carbonates and Neogene molasse-type sediments in the central part (upstream in the Bongka and Balingara rivers) and along the southern coast between Binohoan and Boba villages.

The mountain ranges are characterised by sharp topped peaks with elevations over 700 m above sea level, very rough relief with V-shaped valleys. The Tokala, Batui, and Balingara mountains in the western and middle part, and the Balantak mountain in the eastern part of the East Arm of Sulawesi are the highest ranges in the region. The highest summits include Mt. Tokala (2630m), Mt. Bulu Tumpu (2401m), Mt. Lumut (2284m), Mt. Pasini (2100m) and Mt. Lokai (1600m). The mountain ranges trending SW-NE and are convex towards the north-northwest.

The rivers generally flow in a parallel pattern, and some are dendritic. The principal rivers are the Bongka, Balingara and Binsil rivers which flow northward and then empty into Tomini Bay. The Batui, Toili and Tokala rivers flow southward and empty into the Tolo Gulf. These rivers form V-shaped valleys indicating strong vertical erosion.

1.2. THE OUTLINE OF GEOLOGY OF THE EAST ARM OF SULAWESI

Most geologists and other related earth scientists believe that the geological complexity of the Sulawesi Island and its surroundings is a result of interaction of three major crustal elements, the northward-moving Indo-Australian Plate, the westward-moving Pacific Plate and the south-southeast-moving Eurasian Plate. It is one of the most complicated regions from a plate tectonic point of view.

Sulawesi Island and its surroundings can be tectonically divided into 4 belts, namely , the Western Sulawesi Volcano-Plutonic Belt (WSVPB), the Central Sulawesi Metamorphic Belt (CSMB), the Eastern Sulawesi Ophiolite Belt (ESOB) and the Banggai-Sula (Microcontinent) Platform (Fig. 1.2). Each belt show characteristic lithologic associations and tectonic environments.

The Western Sulawesi Volcano-Plutonic Belt is characterised by Tertiary volcanic and plutonic rocks. Volcanoes have been active since Palaeogene time (Sukamto, 1975a; Simandjuntak et al., 1981; Sukamto and Simandjuntak, 1982), and are still active in the northern part of the belt (i.e. the Minahasa and Sangir-Talaud arcs). The plutonics are granitic rocks of Late Miocene to Pleistocene age (Sukamto, 1975b). The belt also contains thick, flysch-type sediments of Late Cretaceous to Eocene age, and metamorphic rocks which locally occur in association with the intrusive rocks.

The Central Sulawesi Metamorphic Belt (CSMB) consists of a variety of schists, variously in the amphibolite-

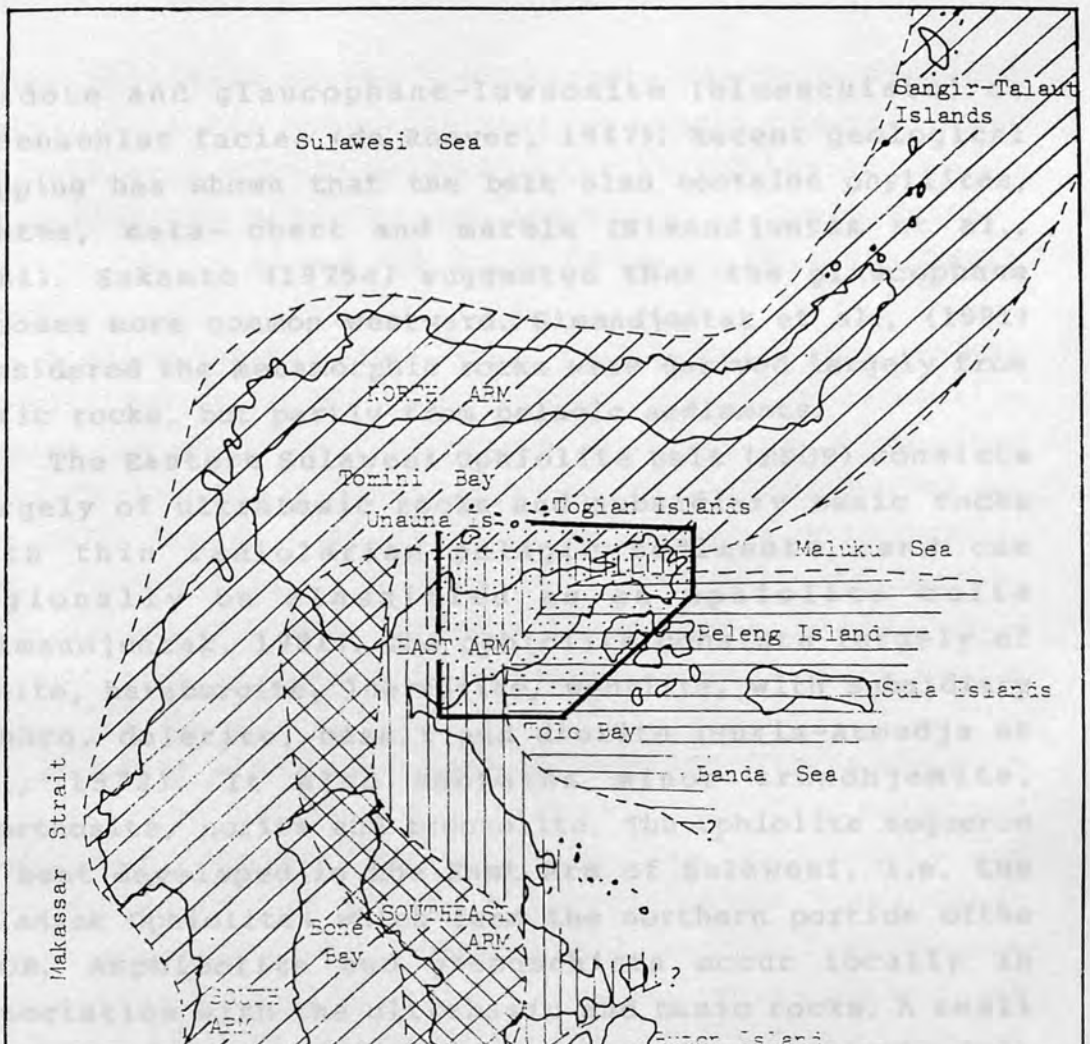


Fig. 1.2 Map showing the four geologic belts in Sulawesi and its surroundings, and the area of this study.

epidote and glaucophane-lawsonite (blueschists) or greenschist facies (de Roever, 1947). Recent geological mapping has shown that the belt also contains phyllites, slates, meta- chert and marble (Simandjuntak et al., 1981). Sukamto (1975a) suggested that the glaucophane becomes more common westward. Simandjuntak et al., (1981) considered the metamorphic rocks were derived largely from mafic rocks, but partly from pelagic sediments.

The Eastern Sulawesi Ophiolite Belt (ESOB) consists largely of ultrabasic rocks and subsidiary basic rocks with thin radiolarian pelagic sediments, and can regionally be classified as an ophiolite suite (Simandjuntak, 1981). The ophiolite consists largely of dunite, harzburgite, lherzolite, wehrlite, with subsidiary gabbro, dolerite, basalt and diorite (Suria-Atmadja et al., 1972). It also contains minor trondhjemite, anorthosite, norite and troctolite. The ophiolite sequence is best developed in the East Arm of Sulawesi, i.e. the Balantak Ophiolites which form the northern portion ofthe ESOB. Amphibolite and greenschists occur locally in association with the ultrabasic and basic rocks. A small exposure of melange also occurs in some parts of the belt (Simandjuntak, 1979). Pelagic sediments, composed of alternating radiolarian calcilutite and chert and minor argillite of Cretaceous age, and are structurally associated with the ophiolite rocks.

The Banggai-Sula Platform (BSP) consists of a metamorphic and igneous basement of Carboniferous to Triassic age and a cover of Mesozoic to Palaeogene continental margin sediments (Sukamto, 1975a; Simandjuntak et al, 1982; Rusmana et al., 1983; Surono and Sukarna, 1985; Supanjono and Haryono, 1985).

In the East Arm of Sulawesi, the Eastern Sulawesi Ophiolite Belt (ESOB) and the Banggai-Sula Platform are juxtaposed and form an imbricated complex. The juxtaposition of these two belts is considered to be due to the collision of the Banggai-Sula Platform with the Eastern Sulawesi Ophiolite Belt in Middle Miocene time. The collision zone is marked by faults and thrusts which are called the Batui Thrust and Balantak Fault System. Melange occurs along the Batui Thrust - Balantak Fault System.

Similar tectonic settings, where a continental margin has collided or is being underthrust beneath a subduction complex, or an island arc system, are reported from other parts of the Banda Arc (Carter et al., 1976; Barber et al., 1977; Bowin et al., 1977, 1980; Barber & Wiryosujono, 1981; Audley-Charles et al., 1981; Cardwell and Isacks, 1978; Johnston, 1981; Silver & Smith, 1983; Smith, 1983) and collision of Luzon arc with Taiwan and SE China (Bowin et al., 1978).

The East Arm of Sulawesi is considered to be the site of collision between the Banggai-Sula Platform and subduction complex. The collision has resulted in the imbrication of the Mesozoic to Palaeogene continental margin sequences with ocean floor material which now constitutes a major ophiolite complex occupying most of the East Arm. Three distinctive types of sedimentary successions occur in the collision zone :

(i) Triassic to Palaeogene continental margin sediments, comprising largely carbonates and subsidiary quartz-rich clastic sediments which contain macroinvertebrate fossils including ammonites, belemnites, gastropods and also microfossils,

(ii) Cretaceous deep sea sediments consisting of chert and calcilutite, rich in radiolaria and associated with the ophiolites and

(iii) the overlying Neogene coarse clastic sediments and volcanogenic turbidites.

There is a transition in the composition of components forming the Neogene coarse clastic molasse-type sediments. The components are dominated by volcanics and metamorphic rocks in western part of the Central Sulawesi, and are dominated by metamorphic and ophiolitic rocks and subsidiary sedimentary rocks to the east of Central Sulawesi (Simandjuntak et al., 1981, 1982; Surono et al., 1984; Rusmana et al., 1984). In the East Arm of Sulawesi, the components are dominated by ophiolitic rocks and sedimentary rocks derived from continental margin sequence. Volcanogenic turbidite deposits which unconformably overlie the ophiolite suite occur only in the northern part of the Poh Head.

At least 4 terraces of Quaternary coralline limestone with elevations of approximately 400, 200, 75, and 20 metres above sea level occur on the south coast, near Luwuk. This feature testifies to the recent rapid uplift of the East Arm of Sulawesi. Seismic activity suggests that the collision may still be in progress (McCaffrey et al., 1982).

1.3 PLAN AND PURPOSE OF THE THESIS

The purpose of this thesis is initially to present a more comprehensive analyses of the structural and tectonic development of the East Arm of Sulawesi with regard to nature of the imbricated Mesozoic to Palaeogene continental margin sequence and the ophiolite rocks in the region. Also to examine the palaeogeography through a basin analysis of Mesozoic and Palaeogene sedimentary successions of the Banggai-Sula Platform which are now juxtaposed with the oceanic material in the East Arm of Sulawesi. Stratigraphic units within the East Arm were examined to determine their sedimentology, petrology, biostratigraphy, with the intention of establishing more detailed subdivisions.

The East Arm of Sulawesi, was recently mapped at a scale of 1:250,000, by the Geological Research and Development Centre (GRDC), Geological Survey of Indonesia, 1981 to 1983. Seven geologists, six assistant geologists and five surveyors were involved in this mapping project. The author was assigned as senior geologist and was fully involved in this project. The results have been set forth in the geological reports and maps of the Poso, Batui and Luwuk Quadrangles at 1:250,000 scale. All data collected during the mapping is incorporated in this thesis.

A total of 110 days (September-December 1983 and July 1985) were spent in the field in detailed studies of selected areas. Data were collected by the author from 6 sections in different parts of the East Arm of Sulawesi, they are (A) Lemo Section, (B) Kolo Atas area, (C) Biak-Poh Section, (D) Bombon river Section, Binsil area, and (F) Balantak Section (Fig. 1.3). These sections were chosen on the basis of geological significance and importance of each section. The sections cover all the rock units occurring in the East Arm of Sulawesi. In addition to these criteria, accessibility and degree of exposure was taken into consideration.

Base maps for systematic geological mapping of at

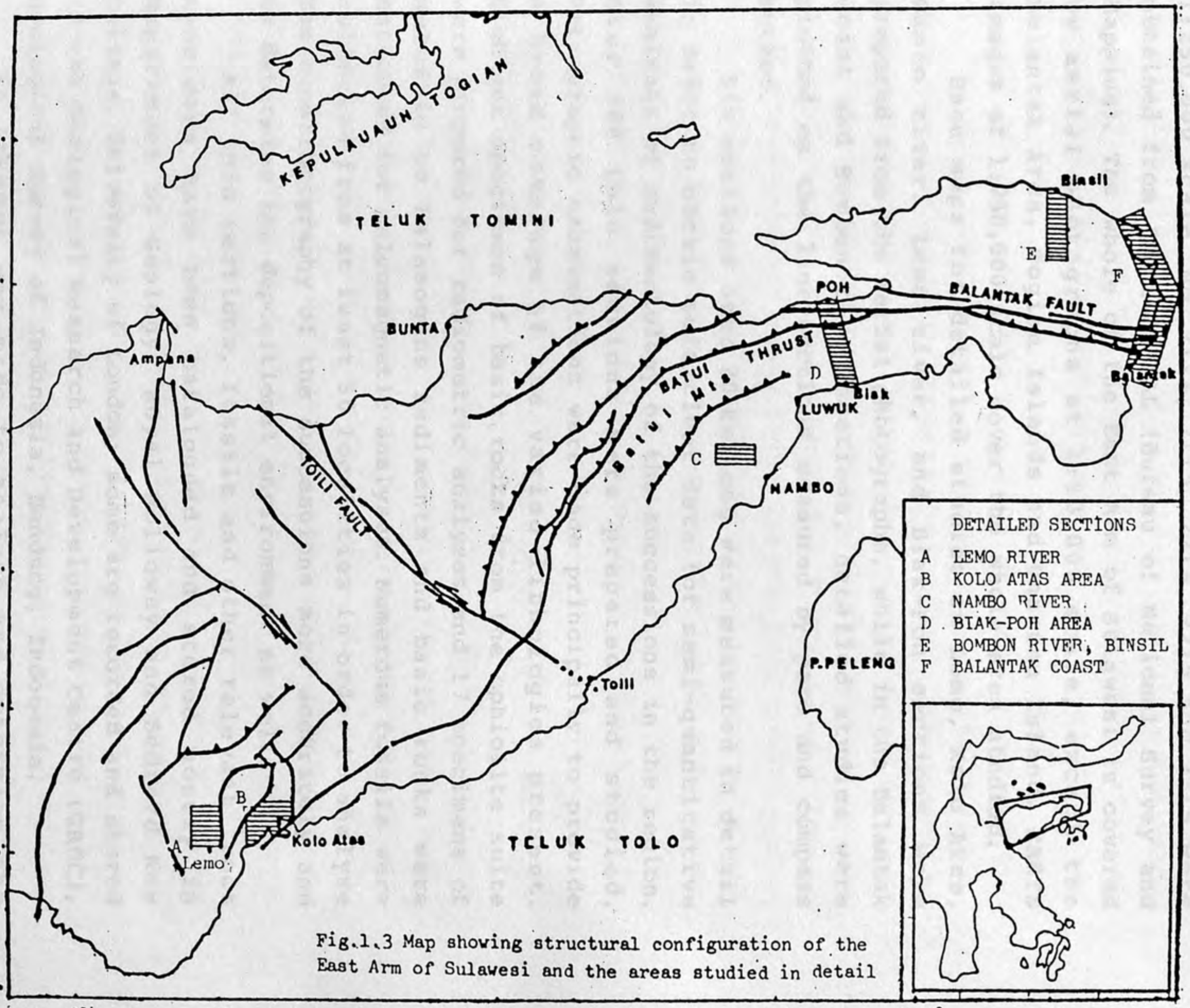


Fig.1.3 Map showing structural configuration of the East Arm of Sulawesi and the areas studied in detail

1:250,000 scale, No. 2114, 2115, 2214, 2215 and 2315 were obtained from BAKOSURTANAL (Bureau of National Survey and Mapping). The whole of the East Arm of Sulawesi is covered by aerial photographs at 1:50,000 scale, except the Balantak Area, Togian Islands and Una-una Island. EARTS images of 1:250,000 scale cover the whole area studied.

Base maps for detailed studies in Lemo, Kolo Atas, Nambo river, Lemo river, and Biak-Poh sections were prepared from the aerial photographs, while in the Balantak coast and Bombon river sections, detailed studies were plotted on the line-sections measured by pace and compass method.

Six sections over 20 km long were measured in detail in order to obtain sufficient data for semi-quantitative analyses of sedimentology of the successions in the region. Over 300 thin sections were prepared and studied. Petrographic examinations were made principally to provide a broad coverage of the varied lithologies present. Eighteen specimens of basic rocks from the ophiolite suite were prepared for radiometric analyses and 17 specimens of Mesozoic to Palaeogene sediments and basic rocks were collected for paleomagnetic analyses. Numerous fossils were collected from at least 50 localities in order to analyse the biostratigraphy of the successions more accurately and to determine the depositional environment as well.

All thin sections, fossils and other relevant rock specimens have been catalogued and stored mostly in Department of Geology, Royal Holloway and Bedford New College, University of London; some are recorded and stored in the Geological Research and Development Centre (GRDC), Geological Survey of Indonesia, Bandung, Indonesia.

An attempt was made to analyse and determine the sedimentary mechanism, vertical and lateral relationships and paleocurrent directions of the Neogene volcanogenic turbidites and some of the coarse clastic sediments in order to obtain a better insight and understanding of basin development in Neogene time. This analysis combines with the proximality index (Pl) of Walker (1967) and the ratio of shale to sand of Lovell (1970) in an attempt to reconstruct the palaeogeography of the region during Neogene times.

An attempt has also been made to measure and analyse all the structural elements present, including bedding, folding, faults and thrusts, cleavage, shearing or foliation, fractures and joints. Evidence obtained strongly suggests the occurrence of several hitherto undescribed major structures, namely, the Batui Thrust, the Balantak Fault System and the Toili Fault System. An attempt is also made to analyse the occurrence, geometry and pattern of all mesoscopic structures and their relationship to major structures.

A tectonic synthesis is presented based on the structural and palaeobasin analysis of the region.

1.4 PREVIOUS WORK

The geology and mineral resources of Sulawesi and its surrounding islands have been investigated for many years. At the beginning of this century, several Dutch geologists have visited and made geological investigation in some part of the East Arm of Sulawesi and the surrounding islands. Their works were related mostly to the search for oil and other mineral resources.

Verbeek, in 1908, briefly examined the coastal areas of the East Arm of Sulawesi and studied river gravels in many places. He also made observations on Peleng Island.

In 1910, Wanner discovered and named the Toili Limestones which he tentatively correlated with the Jurassic sediments of Buru on the basis of lithologic similarity. He also studied the Tertiary sedimentary succession in the East Arm of Sulawesi. He correlated the Late Miocene sediments with sediments in the South Arm, which were named the Celebes Molasse by Sarasin and Sarasin problems in the east of Sulawesi (Fig. 1.4).

The Foreign Capital Investments Law promulgated in 1967 has attracted many foreign companies to carry out mineral and petroleum exploration in Sulawesi. P.T. International Nickel Indonesia (INCO) investigated nickel deposits in the eastern part of Sulawesi. P.T. Tropic Endeavour Indonesia is exploring metallic minerals in Block 2, North Arm of Sulawesi. P.T. Aneka Tambang investigated Block 4, Central Sulawesi. Indonesian Gulf Oil Company is investigating the oil potential on-shore and off-shore in Central, South and Southeast of Sulawesi. In 1982, Union Texas (South east Asia) Inc. started to investigate the oil potential in the Tomori Block, on-shore and off-shore of the East Arm of Sulawesi. Husky Oil Company investigated the Banggai region (Banggai Block), but relinguished the concession in 1984.

Since the commencement of the first Five Year Plan (PELITA) of the Indonesian National Development in 1969, the Geological Survey of Indonesia (GSI) has carried out mineral investigations and systematic geological mapping in Sulawesi Island. In 1979, a geological mapping project has been carried out by the Geological Research and Development Centre (GRDC), Geological Survey of Indonesia. By the end of 1986, it is planned that Sulawesi Island will be entirely mapped at 1:250,000 scale. The author has been fully involved in this geological mapping project.

Suria-Atmadja et al.(1972) described the petrology of ultramafic and mafic rocks from central part of the Eastern Sulawesi Ophiolite Belt (ESOB). Ahmad (1976) studied the geology along the Matano Fault in Central Sulawesi. Sukamto (1975,a,c) briefly discussed the structures of Sulawesi and its surroundings in the light of plate tectonic theory.

Silver et al., in 1976 and 1977, carried out geophysical studies of Maluku Sea Collision zone aboard the R/V Thomas Washington. McCaffrey et al. (1978) carried out earthquake surveys in the Maluku Sea and Sulawesi regions

(1901). He also recognised occurrences of coralline limestone along the south coast of the East Arm, uplifted in very recent times. He was of the opinion that the basic rocks intruded the Palaeogene sediments. Hotz, in 1923, mapped the same area as Wanner (op. cit.) and recognised similar lithologic units, but differed in the interpretation of the structure of the region.

Koolhoven (1930) believed that the gabbros and peridotites were older than the Miocene and that their contact with the Early Miocene limestones was everywhere a thrust fault. His opinion was based on petrographic study of basic rocks near the contacts, which under microscope showed strong pressure effects resulting in the uralitisation of pyroxene and the development of foliated structures.

Von Kutassy (1934) reported the occurrence of 'Permo-Carboniferous' Streptorhynchus, Productus, and Oxytoma from bituminous shale and limestone and Late Triassic Misolia in the limestone from Tokala Mountains. Von Loczy (1934) also found and studied Triassic and Jurassic faunas from Tokala Mountains.

Hopper, in 1940-1941, made a geological reconnaisance in some parts of the East Arm of Sulawesi and Peleng Island. His work was initially related to the investigation of petroleum possibilities in the eastern part of Sulawesi. He briefly described the Mesozoic sediments, but gives more detail on the Tertiary sedimentary successions. Based on the occurrence of large amounts of foraminifera within the Miocene limestones, he was of the opinion that these rocks were the source of the oil seen in the oil seeps occurring in many places in the southern part of the East Arm. He considered that the ultrabasic and basic rocks were intruded into the Early Miocene limestones.

De Roever (1947) studied in more detail the petrology of igneous and metamorphic rocks in eastern Central Sulawesi. Kundig (1956) discussed the geology and ophiolite and found out that the earthquakes occurred at intermediate depths beneath the East Arm of Sulawesi and the eastern Gorontalo Basin. Silver et al., (1978a) made a gravity study in the Central Sulawesi region and suggested that the ophiolite thickened westward.

Simandjuntak (1980) described the occurrence of the Wasuponda Melange in Central Sulawesi and suggested that this melange wedge marked the surface expression of a Cretaceous west-dipping subduction zone. Later, (Simandjuntak, 1981), he discussed the sedimentological aspects of the Mesozoic strata in eastern part of Sulawesi and considered that hydrocarbons were sourced from the shelf sediments of the Banggai-Sula Platform.

Surono (1981) described the molasse sediments in the East Arm of Sulawesi. Smith (1983) studied the sedimentology and tectonics of the Miocene Collision Complex and the overlying later orogenic clastic strata in Buton Island, Southeast Sulawesi, and considered that collision in Buton is older than collision in the East Arm of Sulawesi.

The tectonic evolution of the eastern part of Indonesia, which includes the present area, has been discussed and described on the basis of plate tectonic theory, by Katili (1970, 1971, 1972, 1974, 1975, 1978, 1984), Hamilton (1973, 1977, 1978, 1979). Audley-Charles et al. (1972), Gribi (1973), Carter et al. (1976), Barber and Audley-Charles (1976), Katili & Hartono (1983), Katili & Asikin (1985). The Geological Research and Development Centre (GRDC) published a special publication (Barber and Wiryosujono, 1981) which contains many ideas and findings developed during the last decade from land geological and marine geological and geophysical studies in Eastern Indonesia. Sukamto and Simandjuntak (1982) discussed the relationship between the geologic provinces of Western Sulawesi, eastern Sulawesi and Banggai-Sula from a sedimentological point of view.

1.5 TERMINOLOGY AND CLASSIFICATION

1.5.1 Lithological Terminology

The grade scale of Wentworth (1922) is taken as the basis for subdivision of clastic rocks into various size grades (Table 1.1).

The classification of arenites by Dott (1961), McBride (1963), Dickinson (1970), Folk (1974) and Pettijohn (1975), (Fig. 1.4) and of mudstone by Pettijohn (1975) and the American Geological Institute (1960) are used in this study. The term argillite is used in the sense of Twenhofel (1937), vide Pettijohn (1975) and the American Geological Institute (1960) for incipiently metamorphosed mudstone or lutite. Calcarenite is used for a detrital limestone consisting of grains of sand size, and if the fragments are over 2 mm in diameter, the term calcirudite is applied to the rocks (Pettijohn, 1975). Chert is used in the sense of Pettijohn (1975) and the term cherty-argillite is applied to dense argillite with tough splintery to conchoidal fracture.

A scale of textural maturity proposed by Folk (1962) is used in this study.

The carbonate rocks are classified in the sense of Folk (1959, 1960) and Dunham (1962), and the descriptive names of limestone proposed by Blatt (1982) are used in this study (Table 1.2).

The division of pyroclastics into crystal, vitric and lithic tuffs or combination of these three variants, following Williams, Turner and Gilbert (1954) and Pettijohn et al. (1972) has been employed in this study (Fig. 1.5).

The term ophiolite is used following the definition established by the Penrose Field Conference (1972).

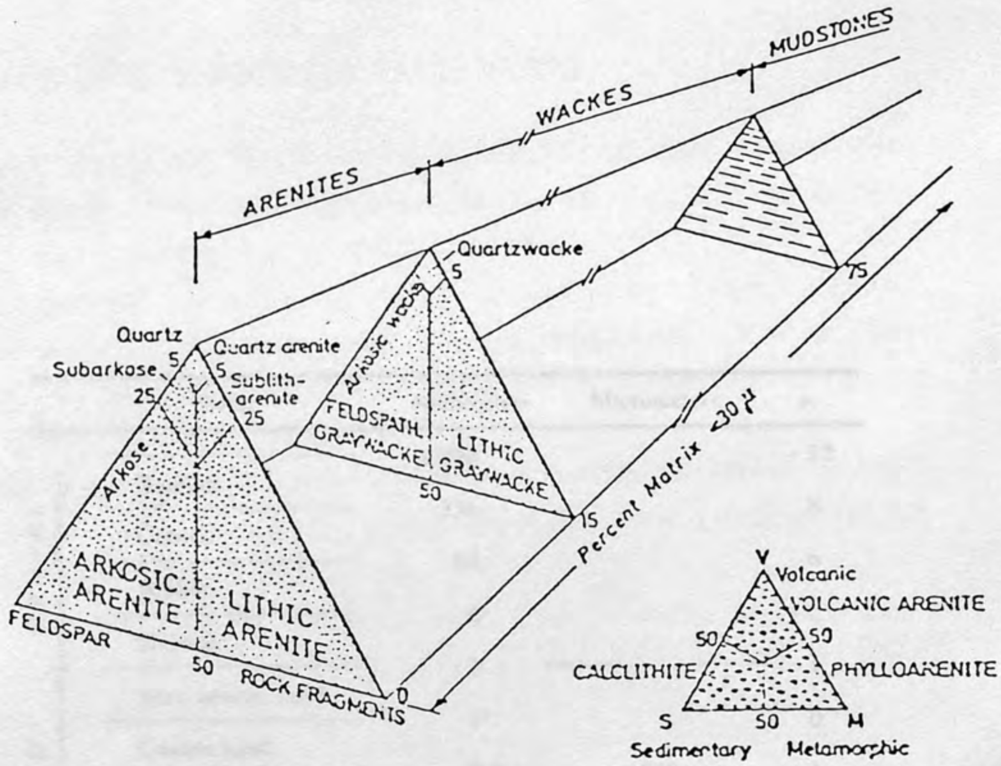


Fig.1.4 Classification of terrigenous sandstones. (after Dott,1964 and modified by Pettijohn et al.,1972).

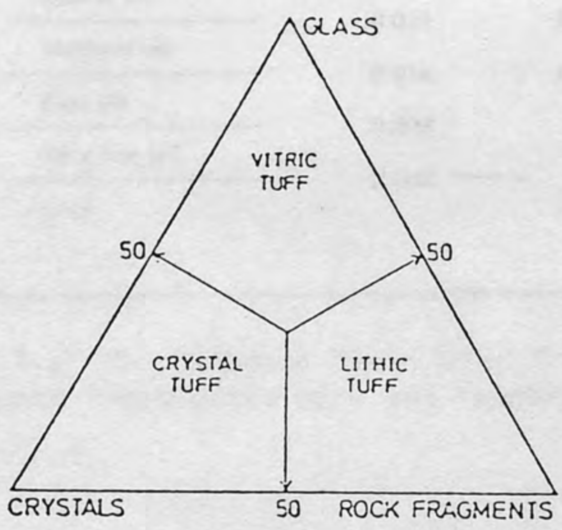


Fig.1.5 Classification of pyroclastic rocks. (after Pettijohn et al, 1972).

	Name	Millimeters	Micrometers	φ
		4,096		−12
GRAVEL	Boulder			
		256		8
GRAVEL	Cobble			
		64		−6
GRAVEL	Pebble			
		4		−2
GRAVEL	Granule			
		2		−1
SAND	Very coarse sand			
		1		0
SAND	Coarse sand			
		0.5	500	1
SAND	Medium sand			
		0.25	250	2
SAND	Fine sand			
		0.125	125	3
SAND	Very fine sand			
		0.062	62	4
MUD	Coarse silt			
		0.031	31	5
MUD	Medium silt			
		0.016	16	6
MUD	Fine silt			
		0.008	8	7
MUD	Very fine silt			
		0.004	4	8
MUD	Clay			

Table 1.1 The Standard Grain Size Scale for clastic sediments (Wentworth, 1924 and Krumbein, 1934)

1.5.2 Sedimentary Structures Terminology

On the basis of structures present in the formation, McKee and Weir (1953) recognised three basically different groups of term (Fig. 1. ; Table 1.3) :

a. **Qualitative terms:** such as stratification, cross-stratification, stratum and cross-stratum, which are concerned with the attitude and relationship of rock layers without regard to scale.

b. **Quantitative terms:** such as thick-bedded, thin-bedded and laminated which are concerned with the thickness of the stratification of layering.

c. **Quantitative terms:** such as massive, slabby and flaggy, which are concerned with thickness of splitting within stratified units.

Stratification is the general term for layering in rocks and it refers both to the process of stratifying and to the state of being stratified. There is no implication concerning the thickness of individual layer involved.

Stratum is a single layer of homogeneous or gradational lithology, deposited parallel to the original dip of the formation. It is separated from adjacent strata or cross-strata by surfaces of erosion; non-deposition or abrupt change in character. Stratum is not synonymous with the terms bed or lamination, but includes both. The term bed and lamination carry definite thickness connotation.

Cross-Stratification is the arrangement of layers at one or more angles to the dip of the formation. A cross-stratified unit is one with layers deposited at an angle to the original dip of the formation. The term cross-bedded and cross-lamination are used synonymously with cross-stratification.

A.

	OVER 2/3 LIME MUD MATRIX				SUBEQUAL SPAR & LIME MUD	OVER 2/3 SPAR CEMENT		
Percent Allochems	0-1 %	1-10 %	10-50%	OVER 50%		SORTING POOR	SORTING GOOD	ROUNDED & ABRADED
Representative Rock Terms	MICRITE & DISMICRITE	FOSSILI-FEROUS MICRITE	SPARSE BIOMICRITE	PACKED BIOMICRITE	POORLY WASHED BIOSPARITE	UNSORTED BIOSPARITE	SORTED BIOSPARITE	ROUNDED BIOSPARITE
1959 Terminology	Micrite & Dismicrite	Fossiliferous Micrite	Biomicrite			Biosparite		
Terrigenous Analogues	Claystone		Sandy Claystone	Clayey or Immature Sandstone		Submature Sandstone	Mature Sandstone	Supermature Sandstone

Folk's (1962) classification of carbonate textures

LIME MUD MATRIX

SPARRY CALCITE CEMEN

B.

Dunham's (1962) classification of carbonate rocks

DEPOSITIONAL TEXTURE RECOGNIZABLE					DEPOSITIONAL TEXTURE NOT RECOGNIZABLE
Original Components Not Bound Together During Deposition				Original components were bound together during deposition... as shown by intergrown skeletal matter, lamination contrary to gravity, or sediment-floored cavities that are roofed over by organic or questionably organic matter and are too large to be interstices.	Crystalline Carbonate
Contains mud (particles of clay and fine silt size)			Lacks mud and is grain-supported		(Subdivide according to classifications designed to bear on physical texture or diagenesis.)
Mud-supported		Grain-supported			
Less than 10 percent grains	More than 10 percent grains				
Mudstone	Wackestone	Packstone	Grainstone	Boundstone	

Table 1.2 Classification of carbonate rocks: A of Folks (1962) and B of Dunham (1962)

A **cross-stratum** is a single layer of homogeneous or gradational lithology deposited at an angle to the original dip of adjacent layers by surfaces of erosion, non-deposition, or an abrupt change in character. Cross-bed or current foreset and cross-laminae are used as synonyms for cross-stratum, but they are restricted to a quantitative meaning. A cross-bed is greater than 1 cm in thickness, and a cross-lamina is 1 cm or less.

The term bed or bedding are applied to any stratum or stratification of thickness greater than 1 cm; laminae and lamination may be applied to any stratum or stratification of 1 cm or less in thickness.

Similarly, cross-bed and cross-bedding may be applied to cross-stratum and cross-stratification of thickness greater than 1 cm, and cross-laminae and cross-lamination may be applied to any cross-stratum and cross-stratification of 1 cm or less in thickness. As adjectives, with specific limits, the following terminology is used in this study : very thick bedded may be applied to strata greater than 120 cm thick, thick-bedded to strata 60 t0 120 cm thick, thin-bedded to strata 5 to 60 cm thick, laminated to strata 2 mm to 1 cm thick, and thinly laminated to strata of 2 mm or less in thickness.

Bedding and lamination should not be confused with terms pertaining to the property of splitting.

1.5.3 Specimen Numbering System

Following the numbering and cataloguing system established by the Geological Research and Development Centre (GRDC), Geological Survey of Indonesia, in this study all rock specimens mentioned in the thesis are numbered in the following way :

83 TO 115.2 or 83 TO 115.B :

83 : The year (i.e. 1983) of collecting the specimen,
TO : The initials of the author,
115 : Location, from which the rock was collected, and
2/B : Indicating that more than one rock sample was taken from this particular locality.

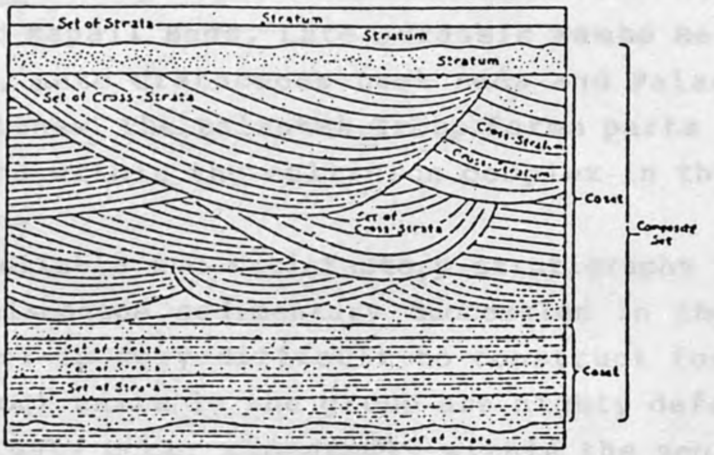


Fig.1.6 Terminology of Stratified and Cross-Stratified Units (after McKee, and Weir, 1935).

Terms to describe stratification		Terms to describe cross-stratification		Thickness	Terms to describe splitting property
Very thick-bedded	Beds	Very thickly cross-bedded	Cross-beds	Greater than 120 cm.	Massive
Thick-bedded	Beds	Thickly cross-bedded	Cross-beds	120 cm. (about 4 ft.) to	Blocky
Thin-bedded	Beds	Thinly cross-bedded	Cross-beds	60 cm. (about 2 ft.) to	Slabby
Very thin-bedded	Beds	Very thinly cross-bedded	Cross-beds	5 cm. (about 2 in.) to	Flaggy
Laminated	Laminae	Cross-laminated	Cross-laminae	1 cm. (about ½ in.) to	Shaly (claystone, siltstone) Platy (sandstone, limestone)
Thinly laminated	Laminae	Thinly cross-laminated	Cross-laminae	2 mm. (about .08 in.) or less	Papery

Table 1.3 Comparison of quantitative terms used in describing layered rocks (after McKee and Weir, 1953).

CHAPTER 2

SEDIMENTOLOGY AND TECTONICS OF THE BALANTAK GROUP

2.1 INTRODUCTION

In this study continental margin sediments occurring in the East Arm of Sulawesi are informally named the Balantak Group. The group consists of Late Triassic Lemo Beds, Jurassic Kapali Beds, Late Jurassic Nambo Beds and Sinsidik Beds, Late Cretaceous Luok Beds and Palaeogene Salodik Limestones. The Balantak Group forms parts of the imbricated zone within the collision complex in the East Arm of Sulawesi.

A fully reliable and satisfactory stratigraphy of the Mesozoic to Palaeogene sedimentary succession in the East Arm of Sulawesi is very difficult to construct for many reasons. All rock units in the group are highly deformed; faults and thrusts occur repeatedly within the sequence. These rock units always occur in fault slivers or fault bounded exposures. In some units fossils are very rare and when present, macroinvertebrates especially are poorly preserved or too crushed for generic identification.

Under these circumstances, it is necessary to make a proper correlation or comparison with the better exposed sections on the adjacent areas, in order to understand the stratigraphy of the region. It is also, for this reason and in accordance with the principles of International Stratigraphic Nomenclature, most of the rock units of the Balantak Sequence are designated 'Beds', as they occur in an incomplete succession.

2.2 STRATIGRAPHY, SEDIMENTOLOGY AND PETROLOGY OF THE BALANTAK GROUP

In the following sections, six rock units of the

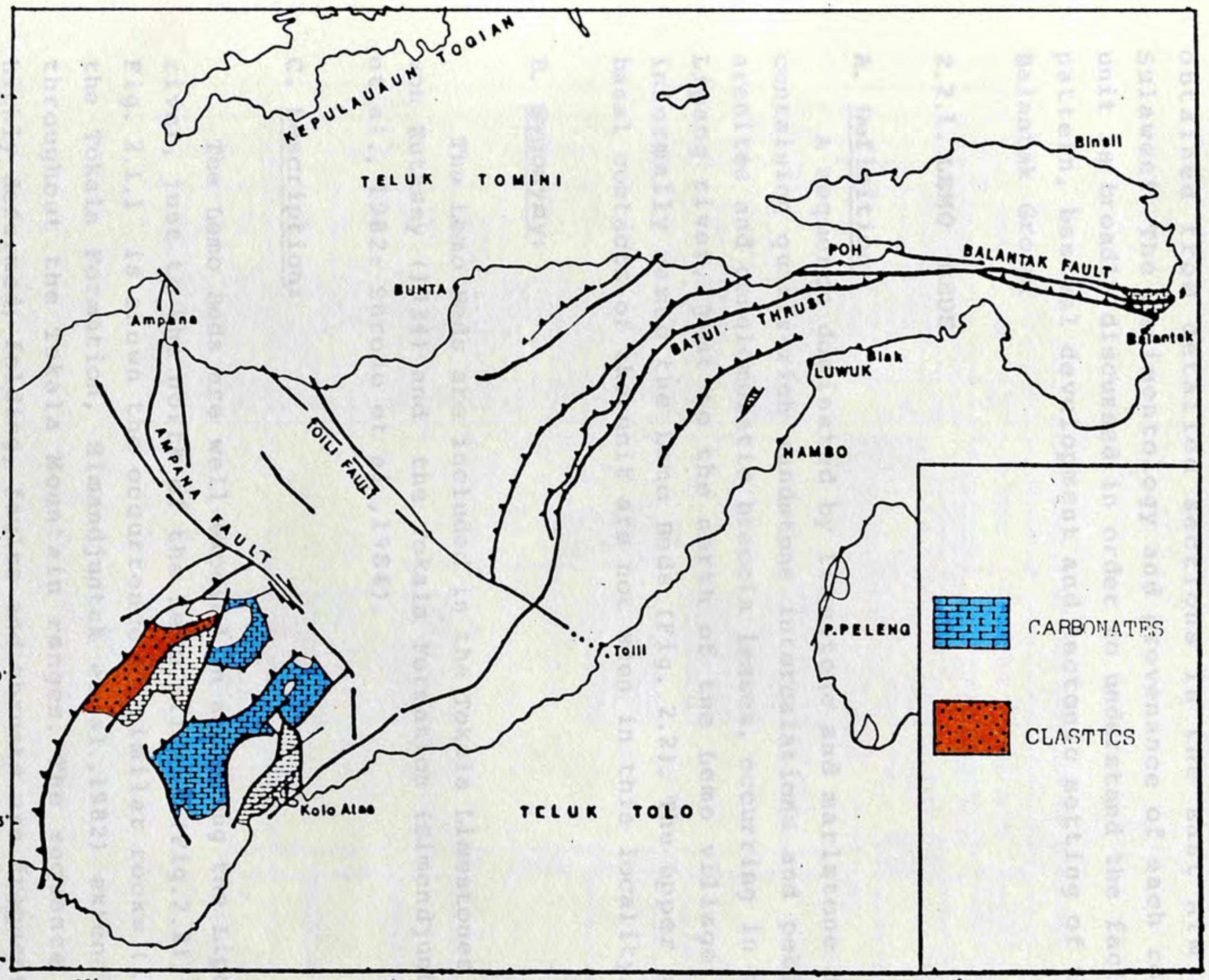


Fig. 2.1.1 Map showing the structural configuration of the East Arm of Sulawesi and the Occurrence of Mesozoic sediments.

Balantak Group ranging in age from Late Triassic to Palaeogene are fully described on the basis of data obtained from detailed sections in the East Arm of Sulawesi. The sedimentology and provenance of each rock unit is broadly discussed in order to understand the facies pattern, basinal development and tectonic setting of the Balantak Group.

2.2.1. LEMO BEDS

A. Definition:

A sequence dominated by limestone and marlstone but containing quartz-rich sandstone intercalations and pebbly arenites and conglomeratic breccia lenses, occurring in the Lipang river, just to the north of the Lemo village is informally named the Lemo Beds (Fig. 2.2). The upper and basal contacts of the unit are not seen in this locality.

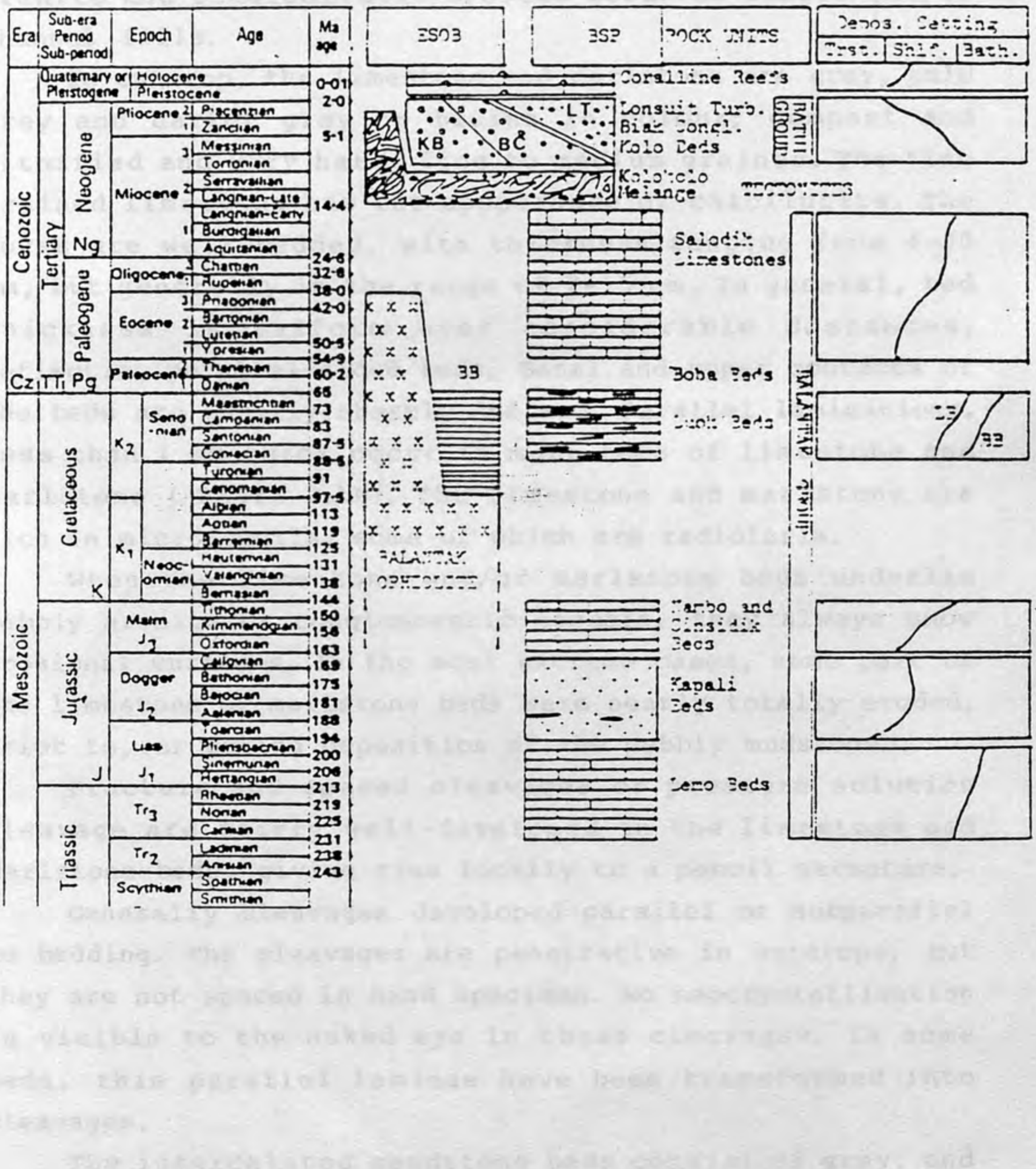
B. Synonymy:

The Lemo Beds are included in the Tokala Limestones of von Kutassy (1934) and the Tokala Formation (Simandjuntak et al., 1982; Surono et al.,1984).

C. Description:

The Lemo Beds are well-exposed in and along the Lipang river, just to the north of the Lemo village (Fig.2.2). In Fig. 2.1.1 is shown the occurrence of similar rocks (i.e. the Tokala Formation, Simandjuntak et al.,1982) extending throughout the Tokala Mountain ranges. The sequence is highly deformed; folding, faults and thrusts are frequently observed in the unit. Bedding invariably dips towards northwest. Sometimes the bedding is up-side down, indicating that the succession has been folded recumbently. Bedding measurements suggest that a regional fold axis

Fig. 2.1.2 STRATIGRAPHY OF THE EAST ARM OF SULAWESI.



trends in E-W direction and plunges, moderately to steeply, towards the southwest (i.e. 230/40 SW).

The unit forms a mountainous topography with steep slopes and karst with dolinas and underground streams. The Lemo Beds consist largely of well bedded limestone and marlstone with alternating sandstone. Subsidiary pebbly arenite and conglomeratic breccia occur as lenses and/or channel fills.

In outcrop, the limestone and marlstone are grey, pale grey and darker grey or bluish in colour; compact and lithified and very hard; fine to medium grained. The fine grained limestone has the appearance of calcilutite. The rocks are well-bedded, with thickness ranging from 4-30 cm, but generally in the range of 7-15 cm. In general, bed thickness is uniform over considerable distances, reflecting parallel-sided beds. Basal and upper contacts of the beds are usually sharply defined. Parallel laminations, less than 1 cm thick occur in most beds of limestone and marlstone (Plate 2.1B). The limestone and marlstone are rich in microfossils, some of which are radiolaria.

When the limestone and/or marlstone beds underlie pebbly arenite or conglomeratic breccia, they always show erosional surfaces. In the most extreme cases, some part of the limestone or marlstone beds were nearly totally eroded, prior to, or during deposition of the pebbly mudstones.

Fracture and spaced cleavages or pressure solution cleavage are fairly well-developed in the limestone and marlstone beds, giving rise locally to a pencil structure.

Generally cleavages developed parallel or subparallel to bedding. The cleavages are penetrative in outcrops, but they are not spaced in hand specimen. No neocrystallisation is visible to the naked eye in these cleavages. In some beds, thin parallel laminae have been transformed into cleavages.

The intercalated sandstone beds consist of grey, and darker grey quartz-rich lithic arenites and subsidiary

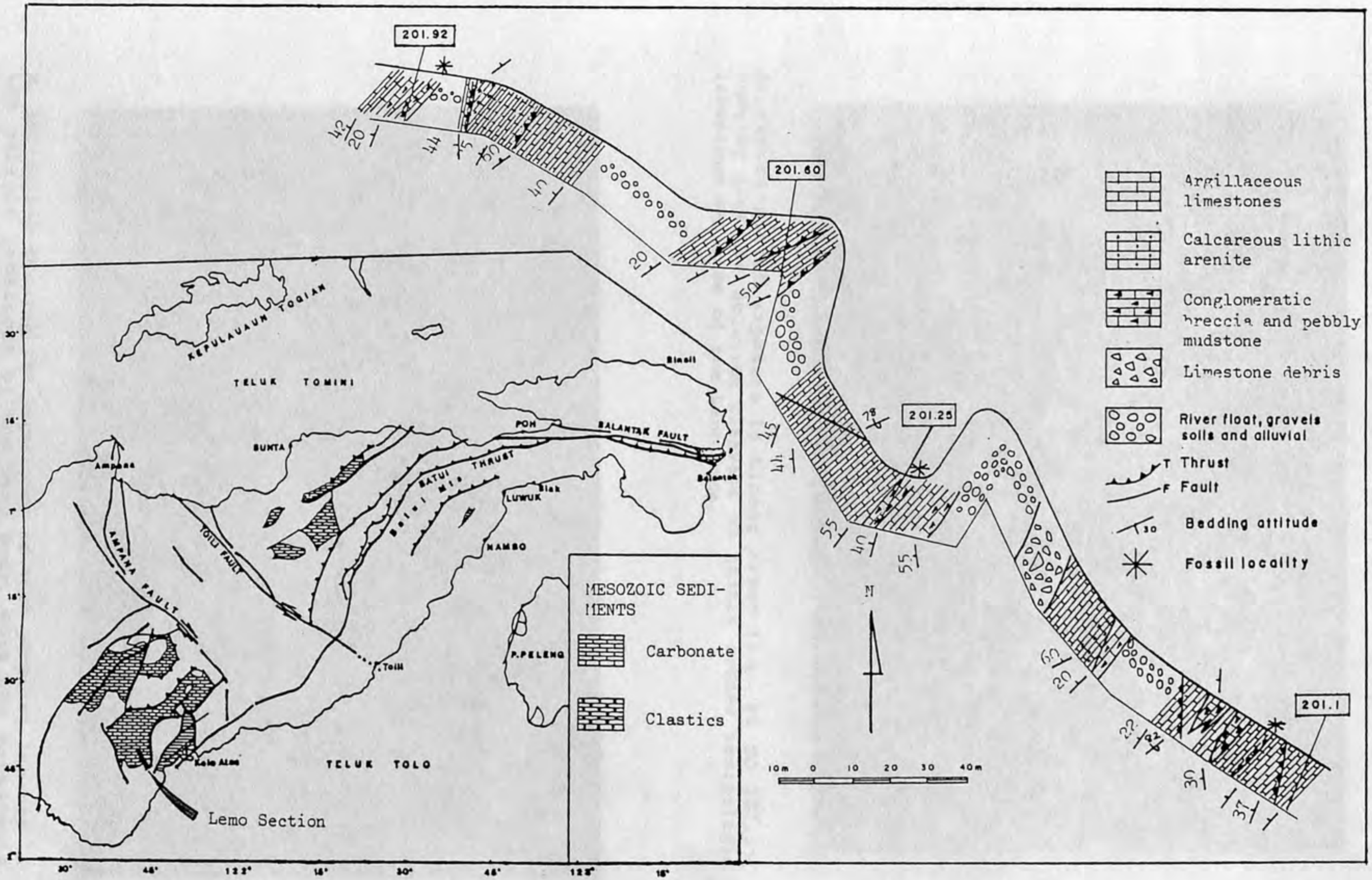


Fig. 2.2 Geological traverse map the Lipang River, Lemo area, showing the occurrence of Lemo Beds.

Plate 2.1 Photographs of the outcrops of the Lemo Beds.

A. Photograph of exposure in Lipang river (i.e. 85 TO 201.1), showing pressure solution cleavage in thinly bedded hemipelagic limestone sequence of the Lemo Beds.

B. Photograph showing the occurrence of thin parallel laminae in the pelagic limestone of the Lemo Beds. Note the mesoscopic thrust just to the left of the camera's cap.

calcarenite, which range in grain size from fine to coarse grained. They are well-bedded, with thickness ranging from 10 to 70 cm. Most beds are uniform in thickness over considerable distances, but a few beds of coarse grained to pebbly arenite, may occur as lenses representing channel fill deposits (Plate 2.1C, D, G).

Some beds of sandstone may show sedimentary features analogous to Bouma's (1962) sequences (Fig. 2.3), but are mostly developed as incomplete sequences, such as Ta; Ta-b; Ta-c; Tb; Tb-c; Tc. In some beds, mud chips of pebble size are present, particularly in the base of the beds. Some of these beds may show grading as indicated by a decrease in both amount and size of the mud chips from base to top. Lumps of coal up to 3 cm long are present in the base of some beds of sandstone. Parallel lamination, less than 2 cm thick, occurs in some beds, and usually forms the Tb interval. Small scale current foreset and convolute lamination are present in some beds usually occurring in the Tc interval. Overall, most of the sandstone beds show very sharp basal and upper contacts. Some of the sandstone beds may show loadcasting features and an erosive basal contact. These features suggest that the arenite and calcarenite beds were deposited by turbidity currents.

Very coarse clastic rocks, occurring as lenses or channel fills within the Lemo Beds, consist largely of matrix-supported pebbly arenites and subsidiary conglomeratic breccia. In outcrop they are dark or darker grey in colour, compact or lithified and very hard; bedded with thickness ranging from 60 cm to nearly 6 metres. Basal and upper contacts are both very sharp. They usually occur on top of eroded surfaces of limestone and/or marlstone beds. The lensoidal and/or channel-filling forms of these rocks are clearly observed in outcrops (Plate 2.1.C, D).

Fragments are typically unsorted, angular to subangular and and some are elongated in shape, with size grades up to nearly 1 m. The fragments consist largely of

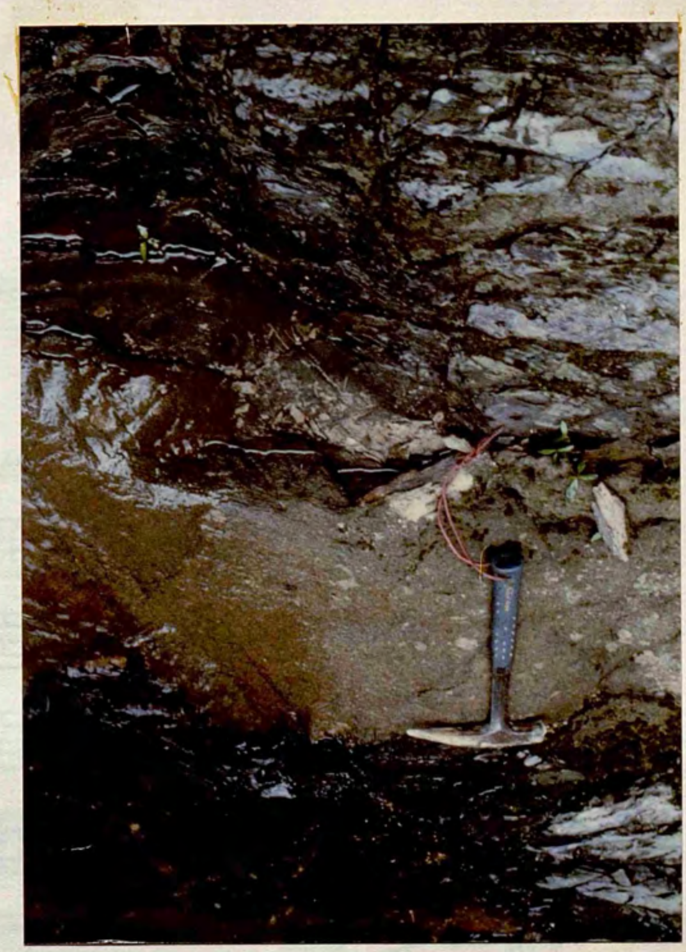


Plate 2.1C Photograph showing a very sharp contact between the hemipelagic limestones and conglomeratic-breccia within the Lemo Beds. The conglomeratic-breccia contains large amounts skeletal debris and significant amounts of terrigenous quartz.

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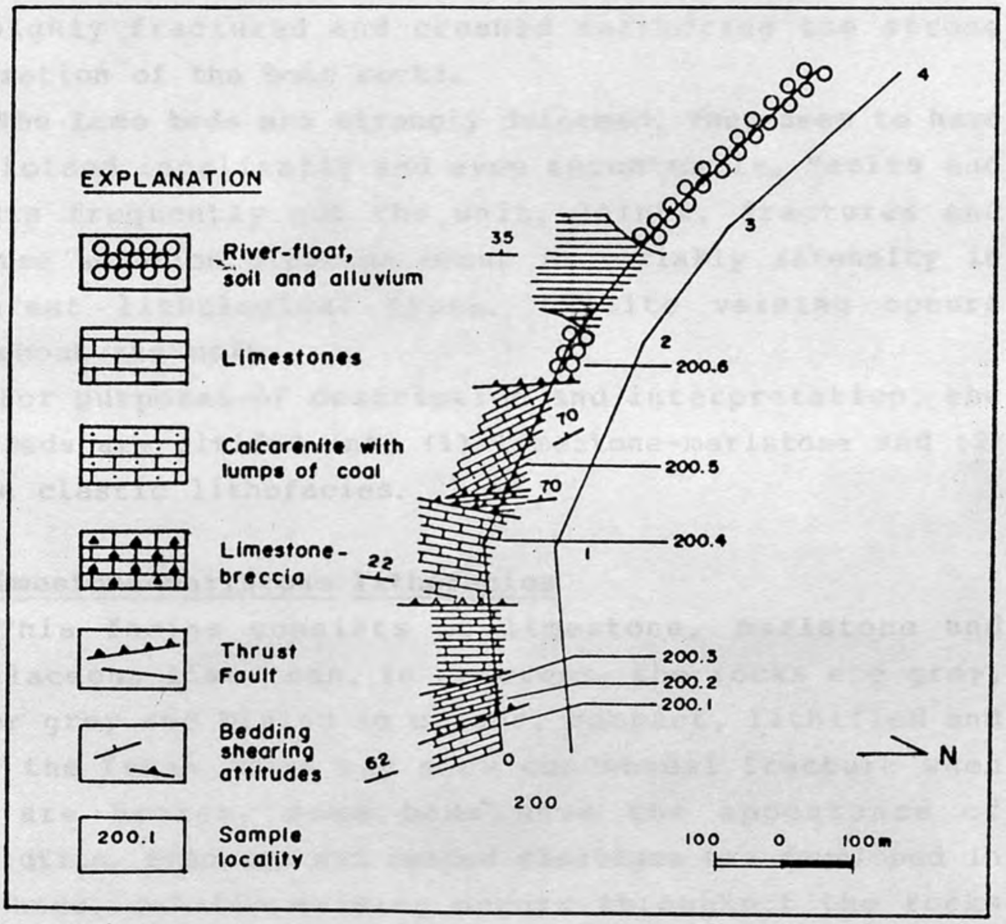


Fig. 2.2B Geological traverse map of Lipang river (83 TO 200.1-6), about 200 m down stream of section shown in Fig. 2.2A.

fossil debris, and subsidiary lithic fragments, including limestone, marlstone and minor metamorphic and volcanic rocks. Fossil fragments include bivalves, ammonites, brachiopods, and crinoid stems. Most of the fossils are too crushed for generic identification. The fossils are very difficult to extract due to the highly cemented nature of the rocks. In outcrops, it is clearly seen that the fossils are highly fractured and crushed reflecting the strong deformation of the host rocks.

The Lemo beds are strongly deformed. They seem to have been folded isoclinally and even recumbently. Faults and thrusts frequently cut the unit. Joints, fractures and pressure solution cleavage occur in variably intensity in different lithological types. Calcite veining occurs throughout the unit.

For purposes of description and interpretation, the Lemo Beds are divided into (1) limestone-marlstone and (2) coarse clastic lithofacies.

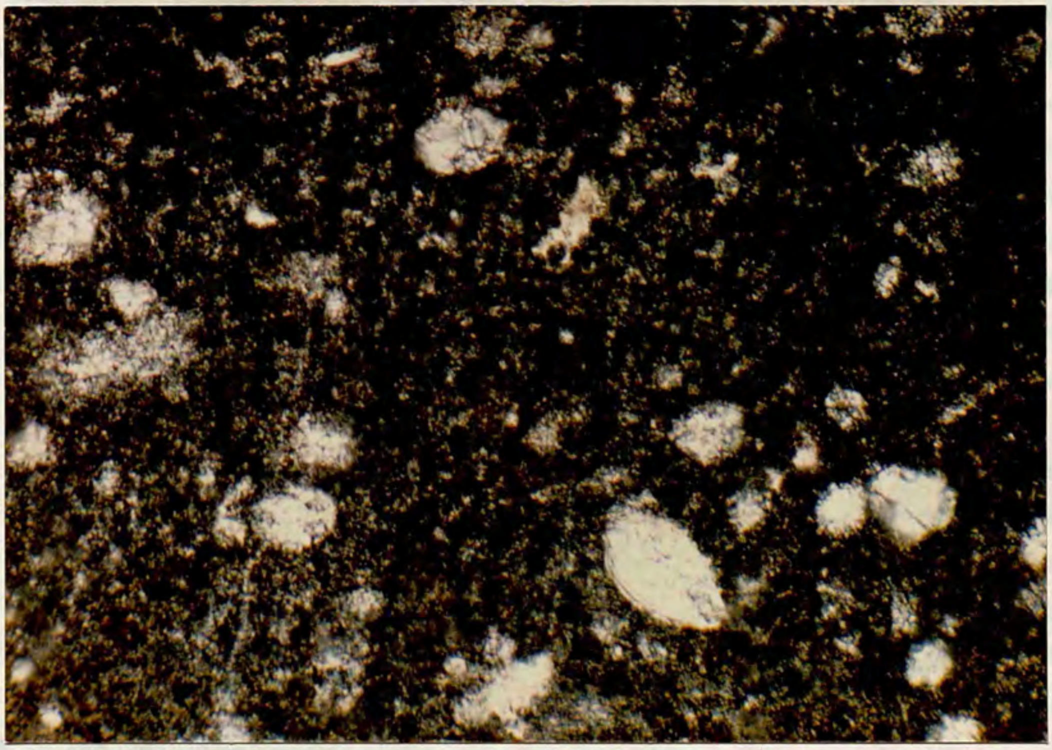
(1) Limestone-marlstone lithofacies

This facies consists of limestone, marlstone and argillaceous limestone. In outcrops, the rocks are grey, darker grey and bluish in colour, compact, lithified and hard; the fresh rocks may show conchoidal fracture when they are broken. Some beds have the appearance of calcilutite. Fracture and spaced cleavages are developed in most beds. Calcite veining occurs throughout the rocks suggesting hydraulic fracturing under condition of high fluid pressure.

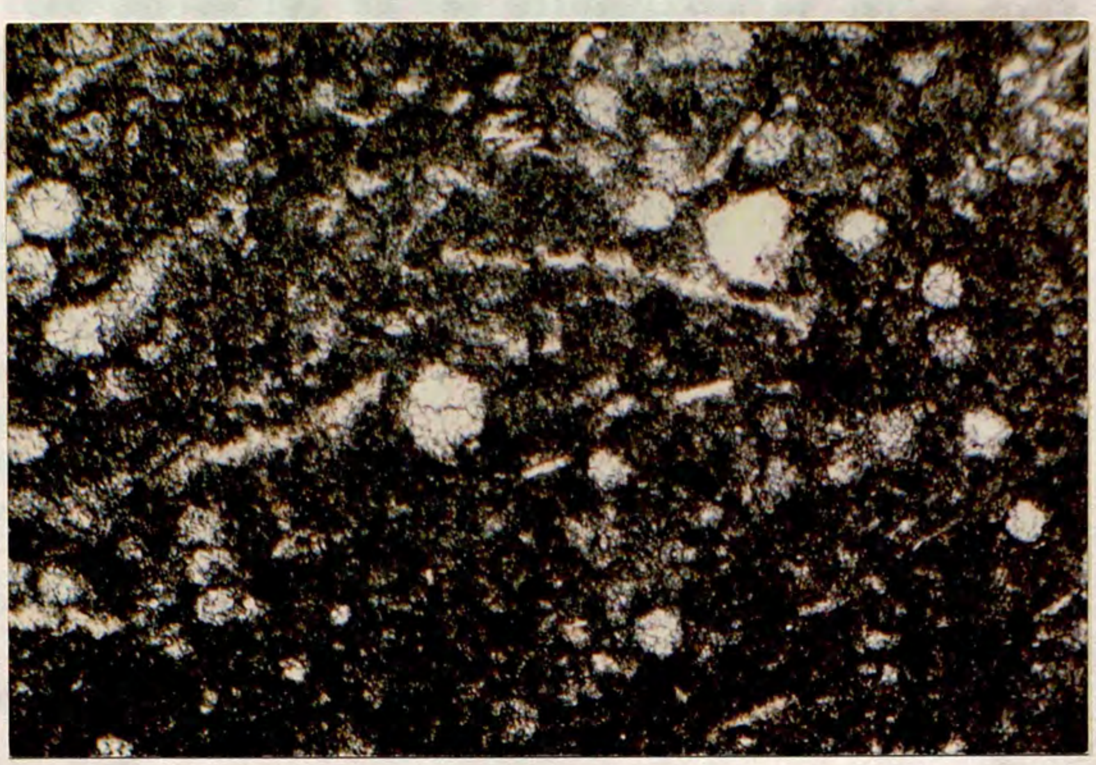
This facies is characterised by presence of significant amounts of microfossils, including nannoplankton and radiolaria. In thin section, the facies consists largely of (a) packstone with subsidiary (b) lime- mudstone, and (c) wackestone.

(a) Packstone Limestone

Plate 2.2. Photomicrographs of limestones of the Lemo Beds



A. Photomicrograph of lime mudstone from the Lemo Beds (85 TO 201.72) showing scattered recrystallised calcispheres of microfossils. Plane polarised light, 125X.



B. Photomicrograph of wackestone in the Lemo Beds (85 TO 201.72A) showing micritised calcispheres of microfossils and abraded skeletal debris. Some of the calcispheres are filled by sparry calcite. Plane polarised light, 125X.

The packstone limestones contain up to 60% fossil fragments, most of which show abraded features, indicating the reworked nature of the debris (Plate 2.2D).Fossil fragments consist largely of calcispheres of microfossils and subsidiary skeletal debris of molluscs and crinoid stems. The microfossils include nannoplankton, possibly the unichambered calcispheres are coccoliths, radiolaria and other planktonic foraminifera. Most of the calcispheres are highly recrystallised and replaced by micrite and some are filled by sparry calcite. The skeletal debris is also micritised (e.g. 85TO201.43).

Detrital quartz of silt size, very angular in shape is present in small amounts (not more than 2%). Some of the larger grains show a wavy extinction, due to plastic deformation. A few grains of prismatic muscovite are also present. Organic matter may be present, but in a very small amount (less than 1%). Iron oxides occur as grain euhedra and in the matrix.

Some of the packstone beds show lamination with thickness less than 2 cm. Laminae are marked microscopically, by the alternation of darker-grey packstone containing fossil fragments, detrital grains of quartz, silt and iron oxides with the pale-grey or yellowish packstone dominated by calcispheres of microfossils and with terrigenous grains either completely absent or present, only in very small amounts (e.g. 85TO201.18).

Development of pressure solution cleavage in the packstone is clearly seen under the microscope, as indicating by the flattened microfossil calcispheres and the occurrence of eyed structure around some calcispheres (Plate 2.2C, D).

(b) Lime mudstone

The lime mudstone consists nearly wholly of micrite matrix with minor and scattered micritised calcispheres of

Plate 2.2C Photomicrograph of packstone from the Lemo Beds (85 TO 200.5) showing the development of pressure solution cleavage. Note the slightly flattened calcispheres of microfossils, some of which show eyed-structure. Plane polarised light, 125X.

Plate 2.2D Photomicrograph of packstone of the Lemo Beds (85 TO 201.11) showing the abraded nature of most skeletal debris, which indicates the reworking of the grains. Micritised microfossils calcispheres are also present. Note the slight flattening of the grains, due to the development of pressure solution cleavage, and the occurrence of calcite veining parallel to and cross-cutting the cleavage. Plane polarised light, 125X.

microfossils; most of them are radiolaria (Plate 2.2A). In laminated mudstone, the laminae contain calcispheres with minor quartz silt and iron oxide euhedra, alternating with pale-grey matrix-supported laminae, or laminae containing rare microfossils (e.g.85TO200.1; 85TO200.3; 85TO201.8; 85TO201.48). In some mudstones the packstone laminae contain up to 40% calcispheres of microfossils, up to 4% very fine grained quartz detritus and up to 7% iron oxides which give rise to the darker colour of the laminae in contrast to the pale-grey micrite-supported laminae (e.g.85TO201.8; 85TO201.58).

In some rocks, calcispheres of radiolaria are slightly elongated or flattened parallel to the pressure solution cleavage (e.g. 85TO200.5; 85TO200.5A;85TO201.7; 85TO201.64). These slides also show micro-calcite veins, most of which are parallel to the cleavage, but some cross-cut the cleavage (e.g.85TO201.13).

Most of the microfossil calcispheres are replaced by micrite, although in some rocks, quite surprisingly, a few calcispheres were siliceous (e.g.85TO201.11). In some rocks, quartz detritus occurs scattered in particular laminae, suggesting that the laminae originated from the settling of the fine-grained particles from quiet water, not from the action of currents.

(c) Wackestone limestone

The wackestone limestones occur interbedded with the lime mudstone and in some turbidite beds they form the Td interval of Bouma's (1962) sequences. In outcrops they are well-bedded marlstone with parallel-sided beds and parallel laminae, not more than 1 cm thick in some beds.

In thin sections, the wackestones are grey, pale-grey or yellowish in colour, and contain up to 40% calcispheres of microfossils including radiolaria, ostracods and spicules (e.g. 85TO201.6; 85TO201.43; 85TO201.47; 85TO201.51; 85TO201.65; 85TO201.68). Other skeletal debris, probably molluscan debris and crinoid ossicles, are present in much less amounts. Most of the skeletal debris shows abraded features indicating that it was reworked (Plate 2.2B). Angular quartz detritus of silt size, is also present in small amounts forming not more than 2% of the rocks (e.g.85TO201.47). Iron oxides occur as grain euhedra and in the matrix, amounting to 3% of the rocks.

The microfossils consist mostly of calcareous calcispheres of planktonic foraminifera, including nannoplankton and radiolaria. Most of the calcispheres are unwalled, and replaced by micrite and some are filled by sparry calcite. These grains are set in a matrix of a mixed lime mud, clay and minor cryptocrystalline quartz and iron oxides. Locally, sparry calcite occurs as cement.

Thin parallel laminae several millimetres thick, seen under the microscope, are defined by the alternation of pale-grey or yellowish wackestone without terrigenous detritus, with the wackestone containing quartz grains and iron oxide euhedra giving rise to the darker colour of the laminae (e.g. 85TO201.50). Some of the laminae have been disrupted by bioturbation or have been transformed into spaced pressure solution cleavage.

In some rocks, the laminae contain angular detrital quartz of silt size, showing slightly imbricated fabric orientation. The long axis of the quartz grains are inclined at 30 degrees to the laminae (e.g.85TO201.50A). These features suggest that detrital quartz was transported and redeposited by infrequent bottom currents, some of which were slightly turbulent. These currents were also responsible for producing the laminae.

Calcite veins on a microscopic scale occur in most of the wackestone beds, indicating the development of hydraulic fracturing under condition of high fluid pressure. Some of the calcispheres of the microfossils are slightly flattened (e.g.85TO201.46).

(ii) Coarse clastic lithofacies

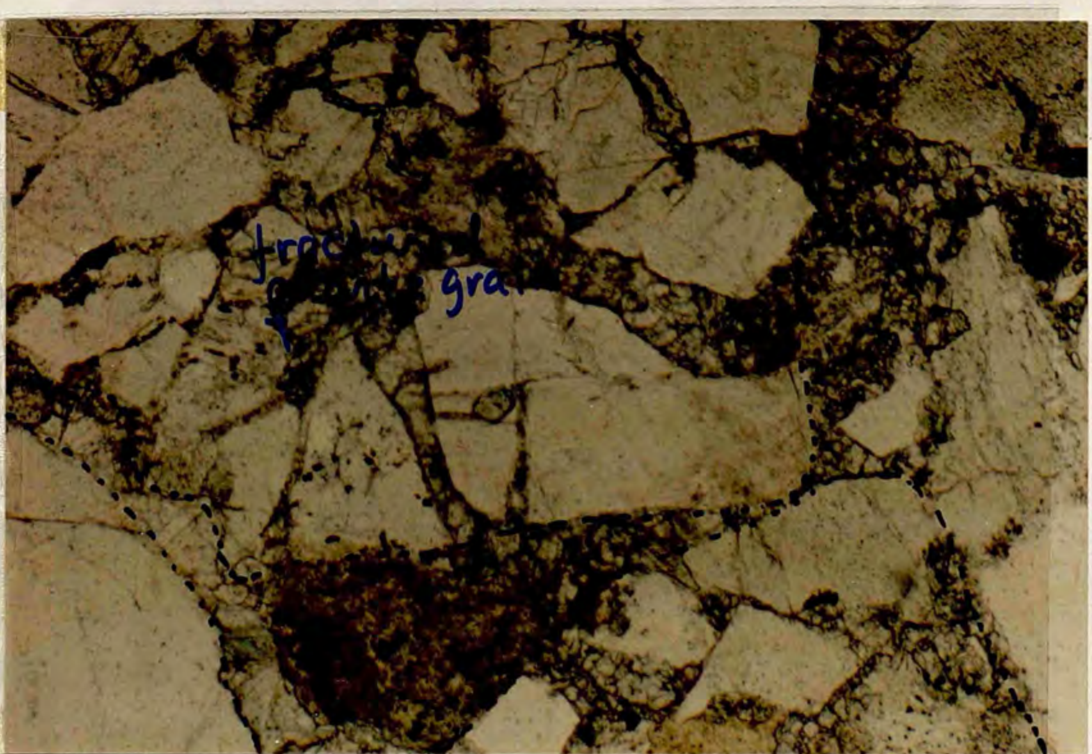
This facies consists of clastic rocks including silty-sands, fine to coarse grained sandstones, pebbly sandstone and conglomeratic breccia. In outcrops, the rocks are pale grey, grey and darker grey and yellowish to brownish on the weathered surfaces; compact or lithified and very hard. The sandstones are characterised by the presence of significant amounts of quartz detritus, which in some rocks, may exceed the calcite grains. The sandstones consist largely of lithic arenite and subsidiary calcarenite.

The conglomeratic breccias contain large amounts of both terrigenous quartz and skeletal fragments including molluscs, brachiopods and coralline algae. Although most of the fossils are fragmented, some nearly complete shells of molluscs are present, including bivalves and ammonites; crinoid ossicles are also present. They are usually well-cemented into a very compact and lithified limestones.

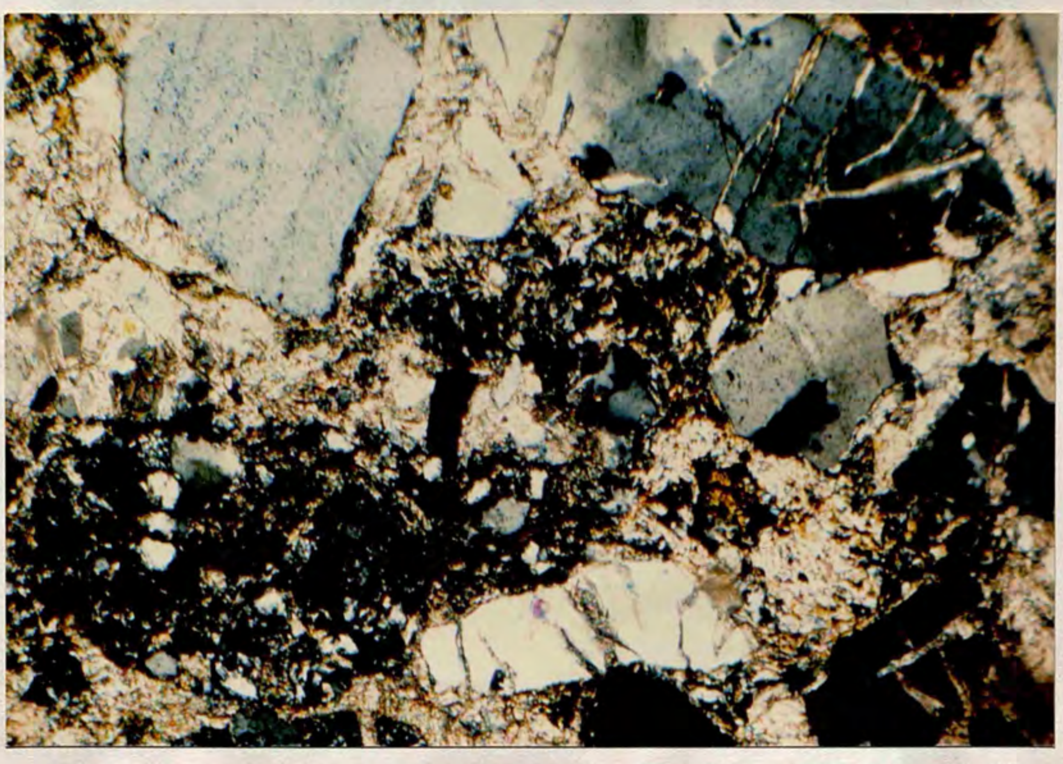
Besides skeletal debris, the rocks also contain lithic fragments which include limestone, marlstone and minor siliceous mudstone, metamorphics and minor volcanic rocks and detrital grains of quartz, feldspar, calcite and minor muscovite and iron oxides, all set in a matrix of a mixed lime mud, clay and iron oxides, and locally sparry calcite. A few lumps of coal up to 2 cm long occur in some beds of pebbly arenite.

The limestone and marlstone fragments are angular and elongate in shape, ranging in size from pebble to nearly 1 m long, but usually in the range of 0.5-7 cm across. In the 85TO201.9, one limestone fragment, nearly 1 m long, shows thin parallel lamination (Plate 2.1G). The lithic fragments are poorly sorted and randomly oriented. The limestone and marlstone fragments, texturally and compositionally, are similar to the limestone-marlstone facies described previously. The fragments are also highly fractured, and some are filled by sparry calcite.

Plate 2.3. Photomicrographs of coarse clastic rocks of the Lemo Beds.



A. Photomicrograph of quartz-rich lithic arenite in the Lemo Beds (85TO201.15) showing highly fractured terrigenous quartz. The original subrounded grains were fragmented and formed several angular chips (grains). Note also the presence of altered volcanic grains (near bottom centre). Plane polarised light, 100X.



B. Photomicrograph of pebbly arenite from the Lemo Beds (85 TO 201.14) showing intensely fractured quartz detritus and altered volcanic fragments (e.g. bottom left corner and in the centre), set in a micritised lime mud matrix. Fractures in the quartz grains are mostly filled by calcite. Crossed polars, 100X.

In thin section, the clastic rocks are grey, pale-grey or yellowish in colour. The clasts consist of limestone, metamorphics, and minor volcanic rocks, angular and elongate in shape, ranging in size from 0.2-4 mm long. The limestone fragments include wackestone, packstone and lime mudstone, most of them contain calcispheres of microfossils and skeletal debris (Plate 2.3C). Most of the fossil grains are micritised and some tests may be filled by sparry calcite. Some of the wackestone and packstone fragments show laminae several millimetres thick (e.g. 85TO201.12). Texturally and compositionally, these limestone fragments are similar to those in the limestone-marlstone facies described previously. These features, combined with the angular shapes of the limestone fragments, suggest that the fragments were transported only a short distance from their source rocks. This is compatible with their occurrence as lenses and/or channel fills.

The metamorphic fragments include schists, gneiss and metaquartzite, rounded to subrounded in shape, ranging in size from 0.2 to 2 mm across. The volcanic fragments are subrounded with size grades up 1.5 mm long, and consist of rhyodacitic rocks which show typical microlitic texture. The volcanic fragments are altered, and form up to 2% of the rocks (e.g.85TO201.90).

Detrital grains present in the arenites and conglomeratic breccia are compositionally similar, consisting largely of quartz, calcite, feldspar and minor iron oxides euhedra and muscovite. They range in size from 0.2 to 4 mm long, angular to subrounded in shape. Some of the quartz and muscovite grains show undulatory extinction and strained features due to plastic deformation. Some of the calcite grains are micritised. Detrital grains may form up to 25% of the rock (e.g. 85TO201.88).

The detrital grains of quartz are mostly monocrystalline, angular in shape, ranging from silt size to 2 mm long. Minor amounts of cryptocrystalline quartz are

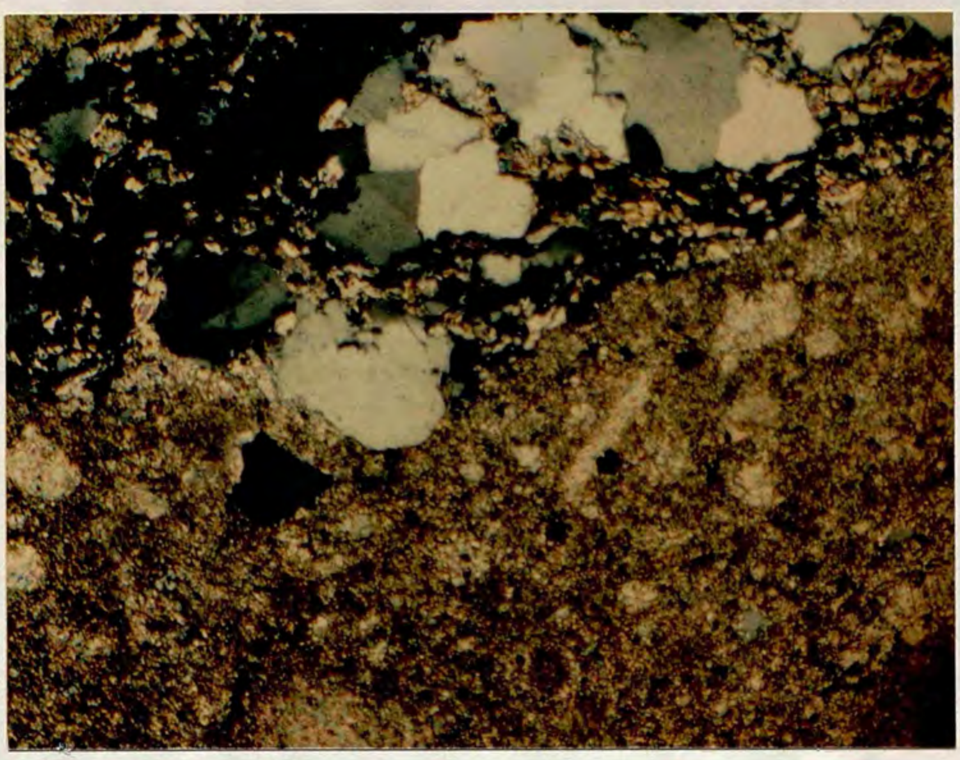


Plate 2.3C Photomicrograph of conglomeratic-breccia from the Lemo Beds (85 TO 201.92) showing fragments of lime mudstone and metamorphics (gneissic, top centre). The lime mudstone fragment contains scattered ghosts of microfossils and skeletal debris, all of which are micritised. Crossed polars,100X.

Plate 2.3D Photomicrograph of pebbly arenite in the Lemo Beds (85 TO 201.14) showing a highly fractured clast of terrigenous quartz and metamorphic fragments. The criss-crossed and randomly oriented fractures in the quartz grains suggests that the fractures were developed prior to deposition of the arenites. Note the subrounded shape of the terrigenous quartz suggesting the polycyclic origin. Crossed polars, 100X.

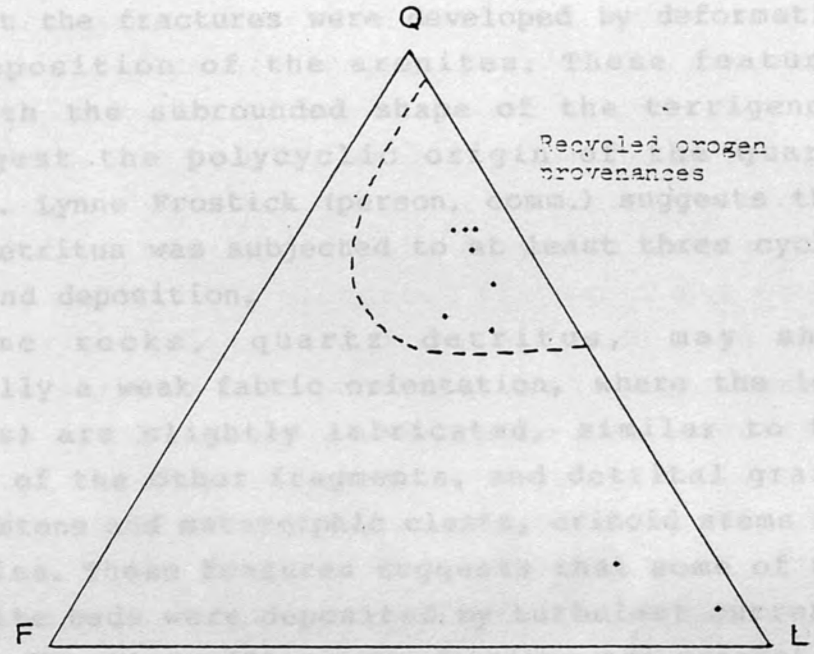


Fig.2.3 Triangular QFL plot (Dickinson,1979) showing that the coarse clastic rocks of the Lemo Beds are derived from recycled orogen provenances.

also present, probably they are derived from recrystallised siliceous mudstone and/or chert. Most of the quartz grains are microscopically fractured, some of the fractures being filled by sparry calcite (Plate 2.3A,B, D). The occurrence of criss-crossed and randomly oriented fractures within the quartz detritus, which do not pass into the matrix, suggests that the fractures were developed by deformation prior to deposition of the arenites. These features together with the subrounded shape of the terrigenous quartz suggest the polycyclic origin of the quartz detritus. Dr. Lynne Frostick (person. comm.) suggests that the quartz detritus was subjected to at least three cycles of erosion and deposition.

In some rocks, quartz detritus, may show microscopically a weak fabric orientation, where the long axes (b-axis) are slightly imbricated, similar to the orientation of the other fragments, and detrital grains such as limestone and metamorphic clasts, crinoid stems and calcite grains. These features suggests that some of the pebbly arenite beds were deposited by turbulent currents (e.g. 85TO201.88). Up to 60% quartz detritus are present in the arenites (e.g. 85 TO 12A).

Detrital feldspar grains are angular and prismatic in shape, but most of their edges are rounded indicating the abraded nature of the grains. K-feldspar and plagioclase are both present, amounting to 2% of the rock (e.g. 85TO201.89). As in the arenites, the K-feldspar includes orthoclase and microcline and the plagioclase is albite and oligoclase, showing carlsbad-albite twinning. The feldspar grains are invariably altered, mostly marginally, and replaced by sericite and carbonates.

Calcite detritus occurs as angular and prismatic grain euhedra, with size grades up to 2 mm long, but generally in the range of 0.4-1 mm. Calcite grains account for up to 5% of the rock (e.g. 85TO201.88). Some of the calcite grains are micritised. Other detrital grains, occurring in very small amounts, include muscovite (less than 1%), and organic matter. Iron oxides occur as grain euhedra of silt size in the matrix.

Fossil fragments include skeletal debris and calcispheres of microfossils, and may constitute up to 45% of the rock (e.g. 85TO201.18). Skeletal debris includes bivalves, gastropods, and crinoid ossicles, which occur in arcuate, oval and elongate shapes. Most of the fragments are micritised, some are filled by sparry calcite. The oval shaped fragments, are usually walled, with the outer rims formed of sparry calcite and the interior filled by micrite or micritised lime mud. Most of the fragments show abraded features, and some of the elongated fragments may show a weak fabric orientation, i.e. parallelism of long axes. These features suggests that these fragments were reworked and deposited by turbulent currents. The calcispheres of microfossils include calcareous nannoplankton and radiolaria. They are micritised and some may be filled by sparry calcite.

The structures, sedimentary features, textures and composition of the pebbly arenites and coarse arenites strongly suggest that they were deposited by debris flows in deep-sea channels. The presence of shell debris suggests derivation from an open shelf environment. The occurrence of microfossils in most of the rocks and the presence of reworked skeletal debris of macrofossils, suggests that the depositional environment was a deep and open sea, i.e. an upper slope environment.

The coarse clastic rocks and the coarser arenites can be interpreted as mass flow deposits, which accumulated in the channels incised within the surrounding fine-grained hemipelagic limestone and marlstone.

D. Stratigraphic relationship between lithofacies

Exposures of the Lemo Beds along the Lipang river indicate that the limestone-marlstone lithofacies are interbedded with the fine to medium grained arenite lithofacies. In some beds of turbidite, the calcareous lithic arenites gradually fine upwards into argillaceous limestone in the upper part of the bed. In most cases, however, a thinly bedded and laminated limestone-marlstone facies occurs in successions of a few metres to hundreds of metres thick, without alternation of the arenite facies. Sedimentary features of the limestone-marlstone facies, including thin and parallel-sided beds, laminated beds, the occurrence of mostly base-cut-out Bouma sequences and the presence of detrital grains of quartz and muscovite in silt size, all suggest that this facies was deposited by bottom currents, some of them may be deposited by low energy turbidity currents.

The coarse clastic rocks and some of the coarser arenite beds, occur in lenses or channel fills. Fragments in these rocks were derived largely from the limestone-marlstone facies, showing that the coarse clastic rocks were accumulated subsequent to deposition and erosion of the limestone-marlstone facies. Turbidity currents and other sediment gravity flows, besides having the capability of carrying a sediment load, also have power to erode previously deposited beds of limestone-marlstone (Middleton and Hampton, 1973).

E. Biostratigraphy

The Lemo Beds contain molluscs, including bivalves, gastropods and crinoids. Most of the fossils are too crushed for generic identification. In the outcrops the fossils, are usually seen to have been cemented firmly within lithified, compact and deformed limestone and marlstone beds.

Most of the limestone beds, especially the wackestone and packstone and some of the lime mudstone, arenites and even the matrix of the pebbly arenite and conglomeratic breccia, contain numerous microfossils, including radiolaria, which are totally recrystallised and deformed, and cannot be identified.

During the mapping of the eastern part of Sulawesi (i.e. Bungku Quadrangle, 1:250,000 scale) we found poorly preserved bivalves of Triassic age (Mr. Fountains, person. comm.) occurring in the limestone and marlstone beds of the Tokala Formation in the Tinala area, south of Bungku township (Simandjuntak et al.,1981).

Von Kutassy (1934) reported the occurrences of 'Permo-Carboniferous' Streptorhynchus, Productus and Oxytoma in bituminous shales and limestones, and Misolia and other Late Triassic fossils in limestones from the Tokala mountains. Kundig (1956) also described the occurrence of Norian Rhynchonella and Misolia in limestones and marlstones in the Tokala Mountains. He also argued, that the 'Permo-Carboniferous' fossils of von Kutassy are Triassic in age. Misolia is a typical Late Triassic fossil found elsewhere in eastern Indonesia.

Von Loczy (1934) also studied similar fossils occurring in the Tokala area, and supposed continuous sedimentation from the Late Palaeozoic up to Late Triassic, although fossils of Early Triassic age have not been found.

On the basis of these faunal assemblages, the age of the Lemo Beds, is considered to be Late Triassic.

F. Discussion and Interpretation

The presence of abundant skeletal debris, which is largely derived from neritic depth in the pebbly arenite, conglomeratic-breccia and in the reworked packstone limestone as well, suggest that the rocks were transported downslope into a deeper water environment, more likely in an upper slope depositional setting. The occurrence of coarse clastic-channel deposits, suggests further, that the the Lemo Beds represent a carbonate sequence of channelised upper slope deposits. Compositionally, the presence of abundant microfossils in the limestone and marlstone and in significant amounts in the arenite and in the matrix of the coarse clastic rocks, suggests that the Lemo Beds were deposited in an open deep sea environment.

Most of the clastic rocks contain variable amounts of detrital grains, including quartz, feldspar, minor muscovite, and lithic fragments, including metamorphic and volcanic rocks. These fragments, undoubtedly have been derived from a metamorphic and plutonic basement complex. Compositionally, these fragments could be derived from the Banggai-Sula Patform, which consists of Permo-Carboniferous metamorphics and Permo-Triassic granitoids in a basement complex, associated with Permo-Triassic volcanic rocks (Sukamto, 1975a; Surono & Sukarna, 1985; Supanjono & Haryono 1985). These fragments were transported together with the macroinvertebrate faunas, downslope from neritic depths into an upper slope environment. The mechanism is essentially mass movement of sediment gravity flows, which carried and transported the sediment load and also eroded the previously deposited sediments. The coarse material was then deposited in channel and/or lensoidal form.

The occurrence of interbedded packstone and wackestone-lime mudstone and the repeatedly occurrence of coarse clastic channel deposits in the Lemo Beds point to deposition of rhythmic alternations in a low energy environment below wave base. The lime mudstone and the wackestone are pelagic and hemipelagic, and the arenites are turbidites derived from mixed carbonate-terrigenous clastic shelf, and the lensoidal and/or channel fill conglomeratic breccia are sediment gravity flow deposits derived from a basement complex and mixed with reworked macroinvertebrate faunas from neritic depth.

Sedimentary features such as coarse arenite turbidites, bioclastic conglomeratic-breccia debris flows and submarine truncation surfaces, which are frequently observed in the Lemo Beds, are typical of carbonate slope sequences occurring in the Tethyan realm, especially during Late Triassic to Early Jurassic time (Wilson, 1969; Evans and Kendal,1977; Evans, et al., 1977).

The juxtaposition of massive, coarse grained arenite and bioclastic conglomeratic-breccia and fine-grained limestone, pelagic or hemipelagic, in a mixed facies is characteristic of an upper slope environment (Walker, 1978).

2.2.2. KAPALI BEDS

A. Definition

A succession dominated by quartz-rich sediments occurring as fault-slivers in the stream course of the Kapali river is informally named the Kapali Beds (Fig. 2.4). The lower and upper boundaries of the unit are not seen in the East Arm of Sulawesi.

B. Description

The Kapali Beds consist largely of quartzose arenite and subsidiary quartz lithic arenite and carbonaceous silty shale. Lumps of coal occur in some beds of arenite. The rocks occur in fault sliver exposures in the stream courses of the Kapali, Gombangi and Pangkape, in the Kolo Atas area. On the basis of lithological similarities, the Kapali Beds are correlated with the Nanaka Formation in the Poso Quadrangle, to the west of Kolo Atas, which is also dominated by quartzose arenite (Simandjuntak et al., 1983; Surono et al., 1984).

The rocks are well-bedded with bed thickness generally ranging from 20 cm to nearly 1 metre, but in the Kapali stream, beds up to nearly 2 metres thick are present. The sandstone beds are massive, hard and structureless, but parallel laminae occur in some thinner beds. Most beds show a sharp base and upper contact. The silty shale beds range in thickness from 3 to 15 cm, some of them contain parallel laminae.

The succession is highly deformed, steeply dipping, highly fractured, jointed and faulted. In the Kolo Atas area, the unit is faulted against melange, ophiolite and chert, and is unconformably overlain by the Neogene Kolo Beds. The thickness of the succession, therefore, is very uncertain. The whole exposed section may not be more than

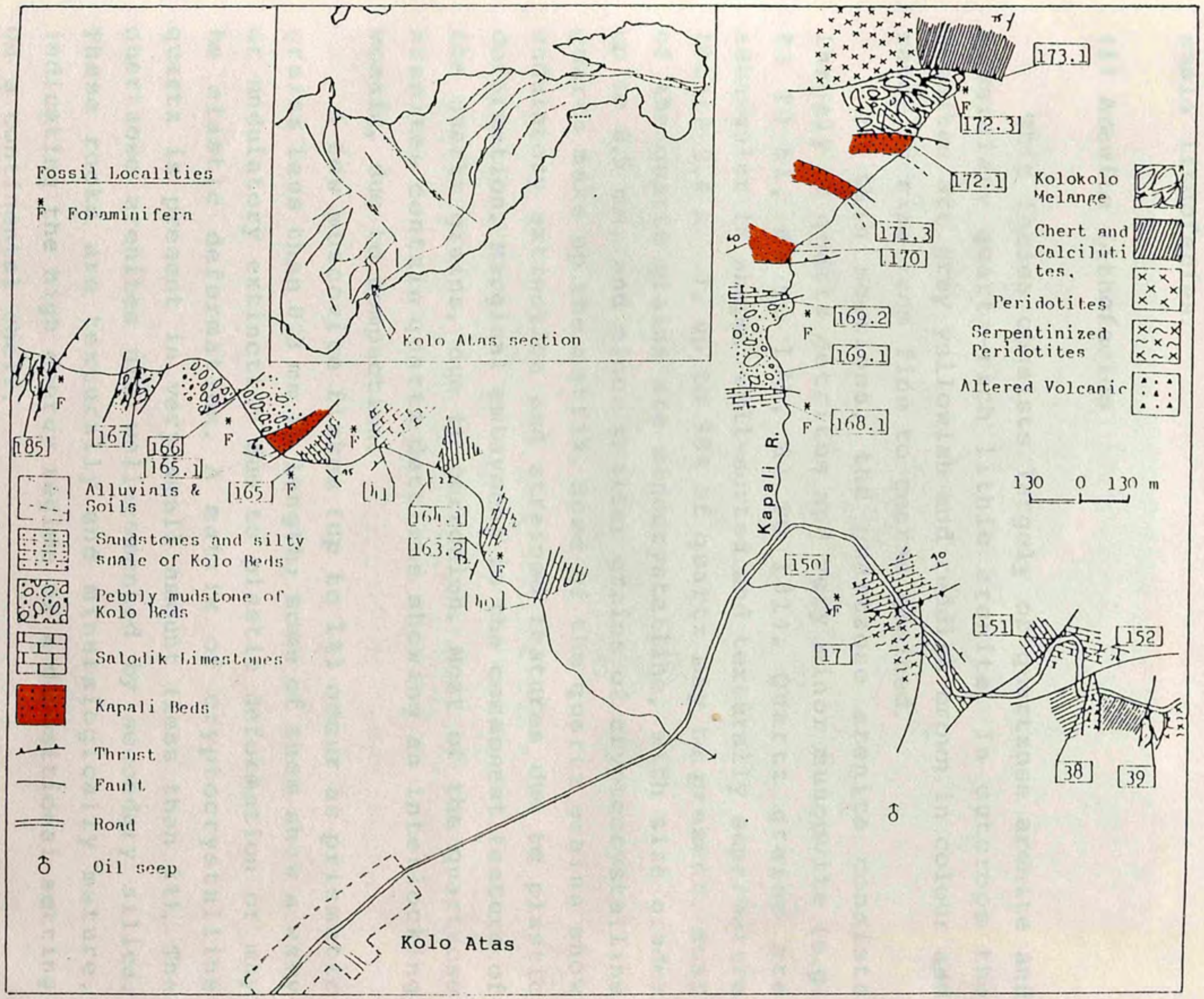


Fig. 2.4 Geological traverse map of Kolo Atas area, showing the occurrence of the Kapali Beds.

50 metres thick. The fault sliver exposures in Kolo Atas area probably represent only a fraction of the original deposit.

For the purposes of description and interpretation, the Kapali Beds are divided into (i) arenite and (ii) silty shale lithofacies.

(i) Arenite Lithofacies

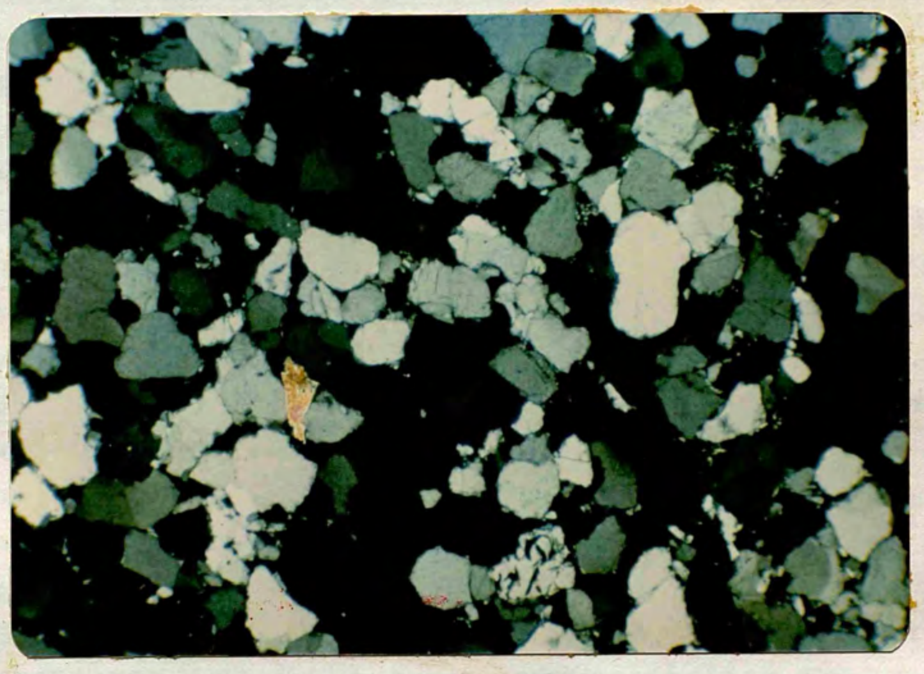
This facies consists largely of quartzose arenite and subsidiary quartz-rich lithic arenite. In outcrops the arenites are grey yellowish and reddish brown in colour and range in size from fine to coarse grained.

In thin sections, the quartzose arenite consists largely of quartz detritus and very minor muscovite (e.g. 83 TO 51, 83 TO 171, 83 TO 181). Quartz grains are subangular in shape, well-sorted and texturally supermature (Plate 2.4 A), up to 98% of quartz may be present. Most of the quartz grains are monocrystalline, with size grades up to 0.5 mm, and minor smaller grains of cryptocrystalline quartz make up the matrix. Some of the quartz grains show undulatory extinction and strained features due to plastic deformation. Marginal embayment is the commonest feature of the quartz grains, due to corrosion. Most of the quartzose arenites contain quartz detritus showing an interlocking mosaic, due to compaction.

A few muscovite flakes (up to 1%) occur as prismatic grains less than 0.3 mm in length; some of them show a wavy or undulatory extinction due to plastic deformation or may be elastic deformation. A matrix of cryptocrystalline quartz is present in very small amount (less than 2%). The quartzose arenites are well-cemented by secondary silica. These rocks are texturally and mineralogically mature, indicating the high energy regime of a depositional setting on a continental shelf.

Lithic arenites are dark and yellowish in colour, fine

Plate 2.4 Photomicrographs of the Kapali Beds

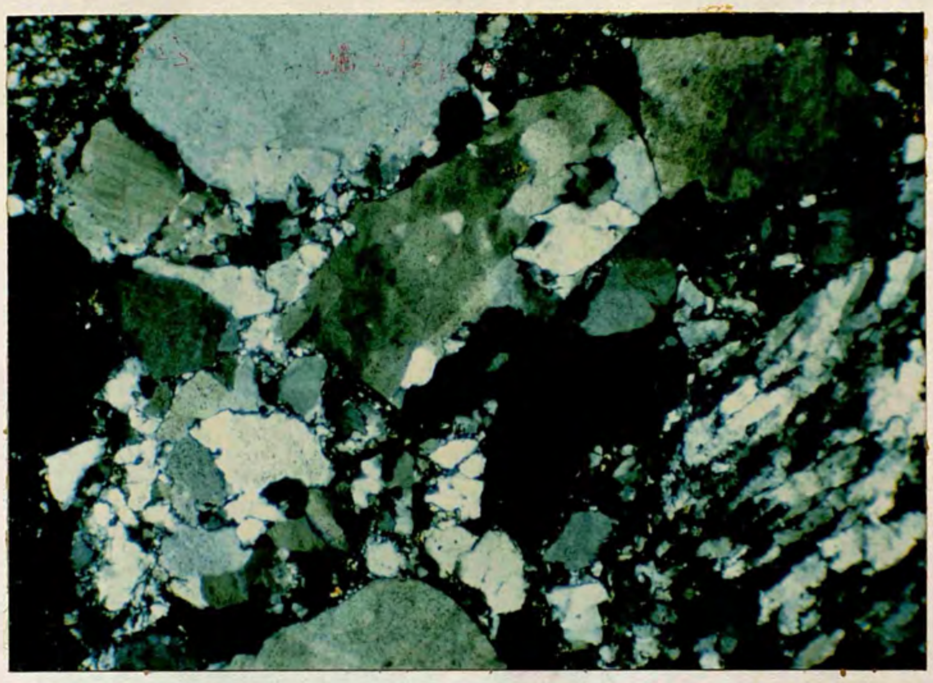


A. Photomicrographs of quartzose arenite from the Kapali Beds showing the interlocking nature of the quartz grains (e.g. 83 TO 171; 83 TO 181). Note the high maturity and well-sorted character of the arenite which reflects deposition under high energy conditions, and a prismatic grain of muscovite near the centre right of photo. Crossed polars, 40X.

to coarse grained, compact and lithified. They are composed of moderately to well-sorted and subrounded to angular grains of quartz detritus, lithic fragments, feldspar detritus and minor chert, muscovite and opaque minerals. In thin sections (e.g. 83 TO 41; 83 TO 171.2) the rocks contain up to 60% quartz detritus, mostly monocrystalline but with minor polycrystalline grains; angular to subangular in shape. The larger grains range in size from 0.2 - 1 mm across, and show graphic texture. Some of them may show undulatory extinction and strained features, due to plastic deformation. The smallest and cryptocrystalline grains occur as matrix; some may be derived from metamorphic rocks and/or recrystallised siliceous mudstone.

Rock fragments form up to 35% of the arenites (e.g. 83 TO 171.2) and consist of angular grains of siliceous mudstone or argillite, slate, schists, gneiss, and andesitic to rhyolitic volcanic rocks, with size grades up to 1 mm across. Some of the siliceous mudstone grains are recrystallised and the volcanics are highly altered. Recrystallised ghosts of radiolarian tests also present in some argillite and chert grains. The volcanic fragments contain plagioclase phenocrysts and some free plagioclase grains are also present. The microlitic texture of these fragments is typical of intermediate volcanic rocks.

Feldspar grains may form up to 5% of the rocks and are mostly altered and replaced by sericite and carbonate or clay. They are mainly K-feldspars, including orthoclase and microcline, characterised by cross-hatched twinning, with minor plagioclase including albite and oligoclase (An 4-11). Feldspar grains are subangular and prismatic in shape, with size grading up to 0.3 mm long. Most of the feldspar grains show abraded features. Muscovite occurs in prismatic shape in very small amounts, and shows a wavy or undulatory extinction.



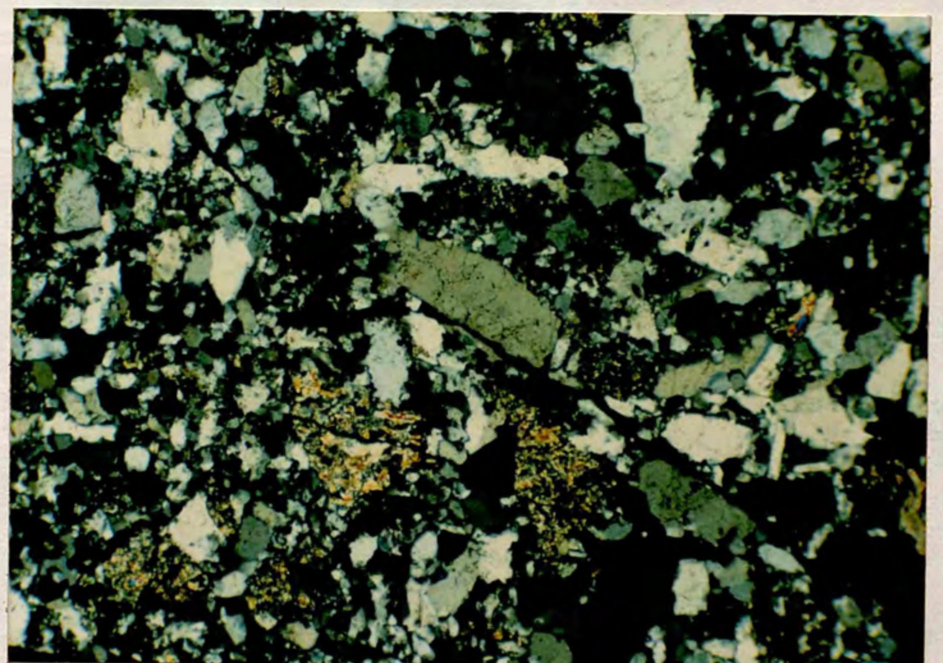


Plate 2.4

B & C Photomicrographs of quartz-rich lithic arenite from the Kapali Beds showing poorly sorted and angular grains terrigenous quartz, and presence of metamorphics (lower left corner of photo B) and altered volcanic fragments (near lower centre of photo C), e.g. 83 TO 171.2. Crossed polars, 40X.

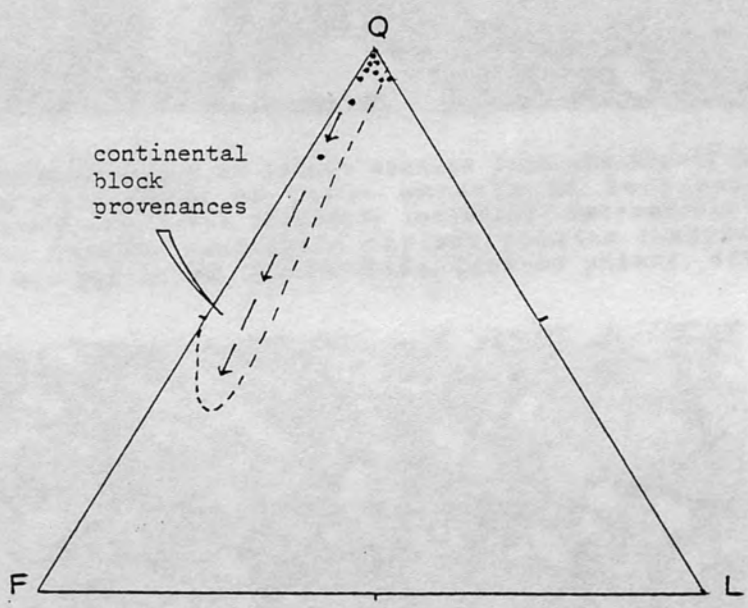


Fig. 2.5 Triangular QFL plot (Dickinson, 1979) showing that the clastic rocks of the Kapali Beds are derived from continental block provenances. Dashed-lined with arrows indicates decreasing maturity and/or stability.

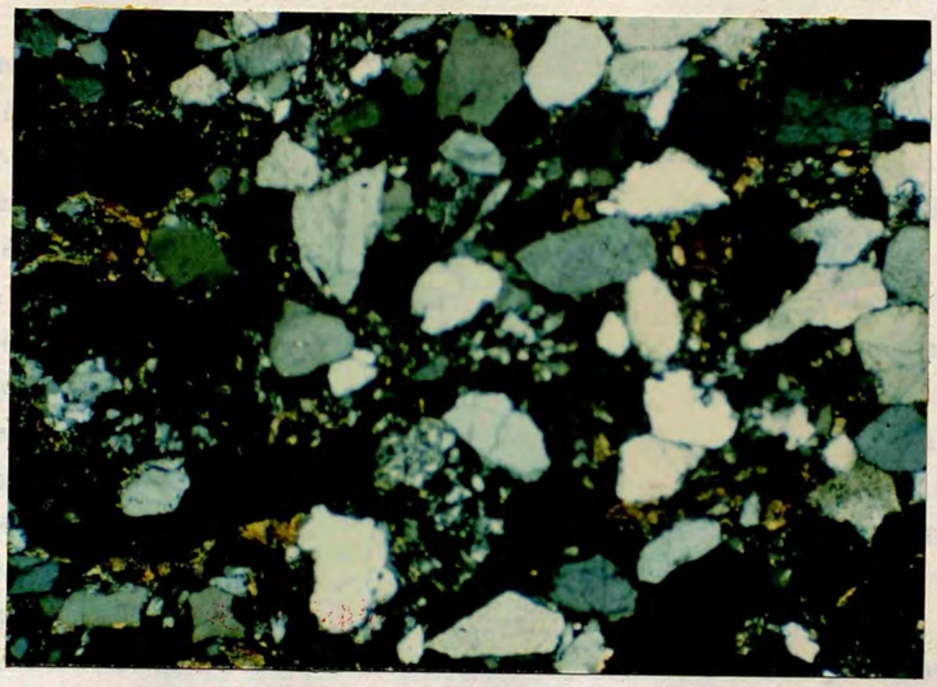


Plate 2.4D Photomicrograph of lithic arenite from the Kapali Beds showing the occurrence of large amounts of terrigenous quartz and subsidiary lithic fragments including metamorphic and volcanic rocks. Note the occurrence of plant remains (very dark), most of which are pyritised (e.g. 83TO165). Crossed polars, 40X.

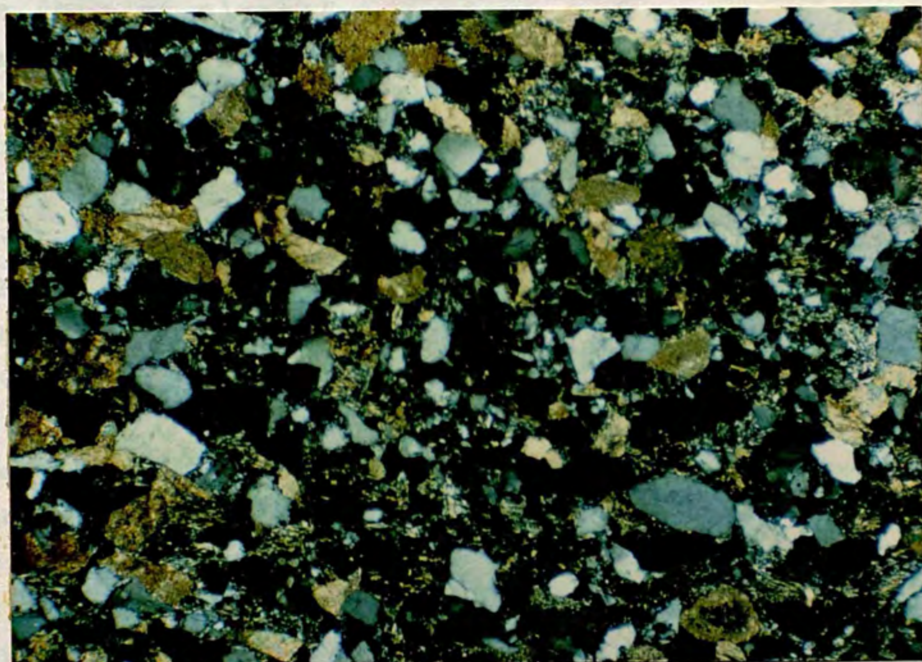


Plate 2.4E Photomicrograph of silty-shale from the Kapali beds showing the occurrence of large amounts of plant remains (very dark) in addition to terrigenous quartz and lithic fragments. Calcite detritus is also present in this rock. (e.g. 83 TO 172.1). Crossed polars, 40X.

(ii) Silty shale Lithofacies

In outcrops this facies consists of thinly bedded, dark and brownish silty shale, which is mostly carbonaceous. In thin sections the rocks consist largely of detrital quartz grains of silt size, and minor feldspar and mica, and significant amounts of plant remains (e.g. 83 TO 171.2). Some of the rocks contain very dark laminae due to presence of plant remains which contrast with the grey, siliceous mud-supported laminae. The laminae range in thickness from a few millimetres to 2 cm. The plant remains are mostly pyritised. The presence of plant remains together with detrital quartz grains in the laminae suggests that they were deposited by bottom currents.

In the stream courses of the Kapali and Gombangi, a rootless coal up to 20 cm thick and up to 50 cm long occurs within the quartzose arenite beds. The lumps of coal are highly deformed, fractured and slickensided, which raise the rank to bright coal (e.g. 83 TO 181.2). The outcrops suggest that the coal fragments were originally deposited in the form of plant fragments, similar to those plant remains occurring in the laminae described previously.

Compositionally, the Kapali Beds, as plotted in the QFL diagram (Fig. 2.5) strongly indicate a continental block provenance (Dickinson, 1979). The quartz, feldspar and muscovite detritus plus rock fragments including siliceous mudstone, slate, schists and gneiss, suggest a source area from a crystalline basement complex. The occurrence of intermediate volcanic fragments suggests that the basement complex wsa associated with volcanic rocks. This description fits the pre-Jurassic of the Banggai-Sula Platform, which consists of Carboniferous (306-305 $\pm$ 6 my) metamorphic rocks including schists, gneiss, amphibolite, phyllite, slate, quarzite, marble and argillite, Permo-Triassic (221-240 $\pm$ 2 my) granitoid rocks and Triassic (210 $\pm$ 25 my) intermediate volcanic rocks (Sukamto, 1975a;

Pigram and Panggabean, 1983; Surono and Sukarna, 1984; Supanjono and Haryono, 1984).

C. Biostratigraphy

No fossils have been collected from the Kapali Beds in Kolo Atas area, and the age of this unit therefore is not definitely known. Hopper (1941), however, described the occurrence of a fairly well-preserved ammonite, Harpoceras cf. toarcense D'Orbigny of Liassic (Early Jurassic) age in a quartz-rich sedimentary succession (his 'Unit B') in the Toili river, to the east of Kolo Atas. Unit B of Hopper (op.cit.) is included in the Lower Jurassic Nanaka Formation, predominantly quartz-rich sediments containing coal fragments (Simandjuntak et a., 1983).

The Kapali Beds, based on lithological and sedimentological similarities, are correlated with the Nanaka Formation and are assigned an Early Jurassic age. On the basis of similar aspects the unit is correlated further with the more firmly established the Bobong Formation dated as Early Jurassic in the Banggai-Sula islands (Supanjono & Haryono, 1985; Surono & Sukarna, 1985) which forms a quartz-rich basal succession unconformably overlying the basement complex (Sato et al., 1978; Westermann et al., 1978; Pigram et al., 1984).

D. Interpretation

The distinctive sedimentological features of the Kapali Beds are the occurrence of massive and structureless quartz-dominated sandstones and the presence of lumps of coal in some beds of the arenite. These features combined with the mature to supermature texture and mineralogy of the sandstones strongly indicate that these rocks were deposited in a high energy regime of an inner shelf depositional setting.

The presence of abundant quartz detritus and the high ratio of K-feldspar to plagioclase feldspar reflect intense weathering on a craton with low relief, and prolonged transport across continental surfaces with low gradients. Furthermore, the typical quartzose arenites, containing minor feldspar, of the Kapali Beds suggest a miogeoclinal wedge depositional setting (Dickinson, 1979). The presence of mature to supermature quartzose arenites or orthoquarzites suggests that the depositional setting is typical of a divergent and passive continental margin (Blatt, 1982).

The Kapali Beds, however, contain quartzose arenites which are gradational to quartz-rich lithic arenites in the upper part. This sequence suggests suites transitional between a cratonic interior and a rifted continental margin (Dickinson, 1979). The composition of the sedimentary sequence may suggest further, that the Kapali Beds mark the first stage in deposition of clastic sediments following separation of the Banggai-Sula Platform from the Australian-Papua New Guinea continental margin at the end of Triassic or in very early Jurassic time. Similar transitional suites are well-described and documented from Mesozoic-Cenozoic sedimentary successions in the Labrador-Greenland margins (Higgs, 1978).

2.2.3 SINSIDIK BEDS

A. Definition

Sinsidik Beds is an informal name assigned to the reddish brown and fossiliferous carbonate dominated sediments, highly deformed and imbricated, exposed in Tanjung Sinsidik, to the north of Balantak village (Fig.2.6). The unit occurs as fault slivers, faulted against ophiolite, and the base and upper portion of the succession are not seen.

B. Description

The Sinsidik Beds consist of reddish brown, fossiliferous limestone interbedded with marlstone and calcarenite in the lower part and argillaceous limestone in the upper portion (Plate 2.5.A). The succession crops out in the fault-sliver exposures along the coast of Tanjung Sinsidik, some 10 km to the north of Balantak village, in Poh Head, East Arm of Sulawesi.

The rocks are well-bedded, but intensely deformed; bedding attitude changes abruptly over very short distances. The rocks appear to have been affected by up-right and tight folds. The exposures are broken by large numbers of faults and thrusts, so that the succession is highly imbricated. Structurally, the Sinsidik Beds occur in a duplex imbricated complex (i.e. the Balantak Duplex discussed in detail in Chapter 3).

The unit is bounded by thrusts against ophiolite at both the northern and southern ends of the exposures. Joints and fractures are the most significant mesoscopic structure occurring in these sediments. Pressure solution cleavage is also weakly developed in some beds of the red limestone and marlstone, particularly within the fault and thrust zones. Most of the fractures are filled by calcite

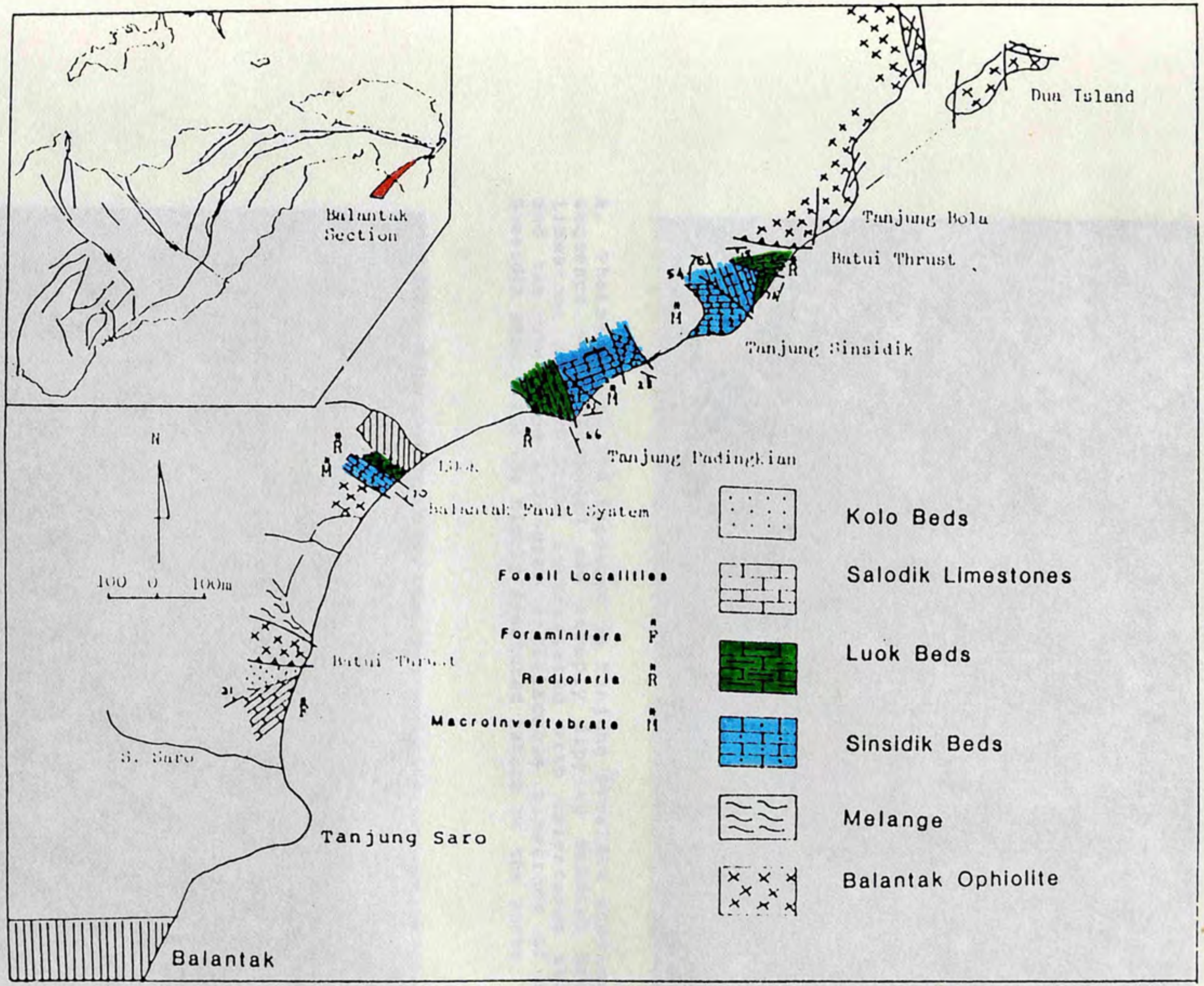


Fig. 2.6.1 Geological traverse map of the Balantak-Tanjung Lonsuit coast, showing the occurrence of imbricated continental margin sediments juxtaposed with the Balantak Ophiolite.

Plate 2.5. Photographs of the outcrops of the Sinsidik Beds

A. Photograph of the exposure on Tanjung Sinsidik showing a sequence of well-bedded and steeply dipping reddish brown limestone and marlstone intercalated with calcareous shale and the overlying pale-grey argillaceous limestone of the Sinsidik Beds. Note the highly fractured nature of the rocks.

B. Photograph of the exposure on Tanjung Sinsidik showing nearly horizontal the rocks, in contrast to those in Plate 2.5.A, C and D.

and some are coated by iron oxides.

For the purposes of description and interpretation, it is convenient to consider the (i) limestone and marlstone, (ii) calcarenite and (iii) argillaceous limestone subfacies of the Sinsidik Beds separately.

(i) Limestone and marlstone Subfacies

The limestone and marlstone are characterised by a reddish-brown colour and contain macroinvertebrates including belemnites, ammonites, bivalves, crinoid ossicles and coralline algae. In outcrops, the rocks are hard, compact and lithified; well-bedded with bed thickness ranging from 5 to 35 cm. Most of the beds are structureless, and the base and upper contacts of each bed are sharply defined. The original internal structures of these rocks have been disrupted by bioturbation, which is frequently observed in these rocks. Most of the larger belemnite guards and ammonite shells are aligned at a low angle to the bedding planes (Plate 2.5.G), suggesting that they have been reworked and deposited by infrequent low velocity bottom currents. Most of the fossils are highly fractured and intensely crushed due to deformation.

In thin section, these rocks consist largely of skeletal fragments which are mostly micritised and/or infilled by sparry calcite and show evidence of abrasion. Skeletal debris makes up 50% of the rocks (e.g. 83 TO 34.1; 83 TO 69.4). Micritised calcispheres of microfossils are also present in some samples.

Additional grain types, present in much smaller size and amounts, include calcite and quartz with size grades up to 0.3 mm. The terrigenous quartz grains are angular and monocrystalline (e.g. 83 TO 34.1). Up to 7% of glauconite pellets of silt size are present (e.g. 83 TO 61), and small phosphatised skeletal grains (possibly fish bones or teeth) are also present in some rocks. Sparry calcite occurs

C

D

Plate 2.5C&D Photographs of the exposures on Tanjung Sinsidik showing highly deformed argillaceous limestone interbedded with calcarenite of the Sinsidik Beds.

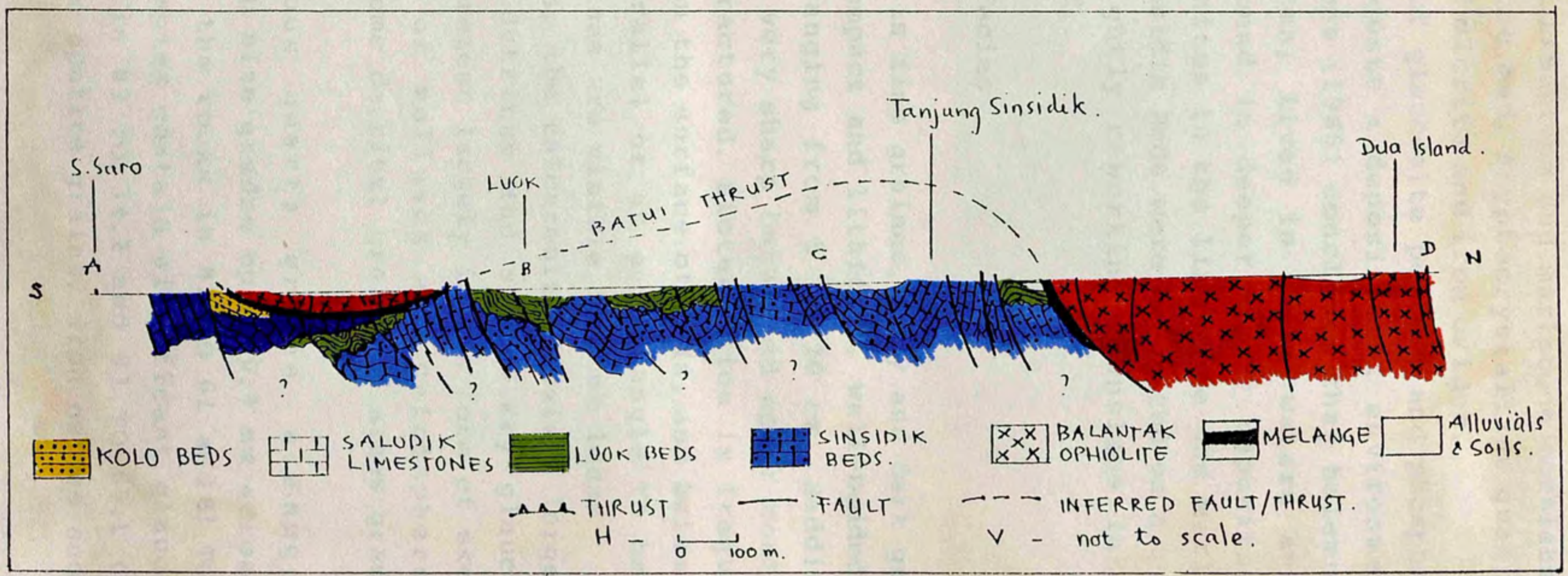


Fig. 2.6.2 Line section from Tanjung Saro to Dua Island, Balantak area, showing the imbricated nature and development of duplex structure of the Mesozoic to Palaeogene continental margin sediments.

within some walled tests of microfossils (e.g. 83 TO 34.5). The matrix of the red limestone and marlstone consists of a varied mixture of lime mud, cryptocrystalline quartz and clay with a cement of micrite and iron oxides.

The occurrence of glauconite pellets and phosphatised skeletal grains suggests a depositional environment of neritic depth. Stevens (1965) concluded that belemnites, though nektonic forms, lived in shelf waters and are therefore rarely found in deeper water deposits. The occurrence of belemnites in the limestone and marlstone suggests that the Sinsidik Beds were deposited on a shallow shelf, probably slightly reworking downslope into the middle or outer shelf.

(ii) Calcarenite Subfacies

The calcarenite is fine grained, grey and dark grey in colour, very hard, compact and lithified, well-bedded with bedding thickness ranging from 5 to 20 cm. Bedding is parallel-sided with very sharp basal and upper contacts; highly jointed or fractured. Bioturbation is frequently observed occurring on the surface of beds, and belemnites usually occur subparallel or at a low angle to bedding planes. Parallel laminae are visible in some beds.

In thin sections, the calcarenite consists largely of calcite and quartz detritus and subsidiary glauconite pellets. Calcite is present largely in the form of skeletal fragments, mainly of molluscs and calcispheres of microfossils, with some detrital grains of sizes grades up to 0.3 mm long.

The terrigenous quartz grains are angular, monocrystalline with size grades up to 0.3 mm across and making up to 25% of the rocks in 83 TO 61 and 83 TO 69.3 (Plate 2.6D). All samples contain significant glauconite pellets, up to 20% in 83 TO 34.2 and 83 TO 69.1 (Plate 2.6A). A few prismatic apatite grains, iron oxides and very

E

F

Plate 2.5E&F Photographs of the outcrops on Tanjung Sinsidik showing the presence of trace fossils, which are mostly of exogenetic type occurring near the surfaces of beds.

minor muscovite are also present. The matrix consists of a mixed lime mud, cryptocrystalline quartz and clay with cement of micrite and iron oxides.

The suite of detrital grains is identical to that of the reddish limestone and marlstone, except for the absence of large shell fragments, and a much higher glauconite content.

The presence of glauconite pellets suggests the depositional setting was in the range of 30 to 1000 metres depth (Blatt, 1982).

(iii) Argillaceous Limestone Subfacies

The argillaceous limestone is very hard, compact and lithified, pale-grey and yellowish brown in colour, well-bedded with bed thickness ranging from 4 t0 20 cm, with sharp bases and upper contacts (Plate 2.5C, D). The outcrops suggest that the thickness of this unit is less than 30 metres.

In thin sections the rocks are yellow brownish wackestone composed largely of skeletal fragments and numerous calcispheres of microfossils, largely of planktonic foraminifera (e.g. 83 TO 60). The skeletal debris consists of fragments of macroinvertebrates including molluscs, coralline algae and crinoid ossicles. Most of the calcispheres have been replaced by micrite and some infilled by sparry calcite.

Scattered fine grained quartz detritus and opaque minerals of silt size are present in small amounts. The matrix is dominated by micrite and minor clays. Micro-fractures are filled by sparry calcite and some are coated by iron oxides.

The original deposit appears to have been formed by a mixture of skeletal debris, calcareous microfossils and fine grained terrigenous quartz, but their original textures have been largely modified by diagenetic

Plate 2.5G Photographs of the exposures on Tanjung Sinsidik showing the occurrence of Late Jurassic <u>Belemnopsis uhligi</u> Stevens on the surface of marly shale bed in the Sinsidik Beds. Note the belemnite is gently inclined to the bedding surface.

Plate 2.5H Photograph of exposure near Luok village showing the print and cast of undetermined ammonite in the silty shale of the Sinsidik Beds.

redistribution of the biogenic carbonate. Some beds of the argillaceous limestone show the features of hard-ground with nodules a darker colour than the host rocks.

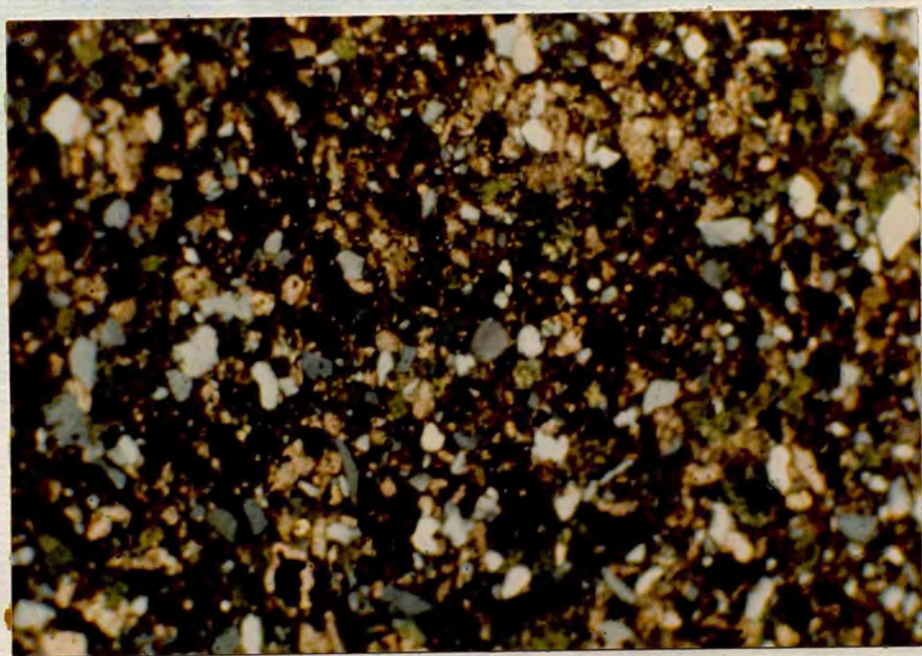
Thin parallel laminae (usually less than 2 cm thick) within the marlstone and calcarenite beds are well-defined by the alternation of grainstone and packstone laminae up to several millimetres thick. The packstone laminae have a matrix of lime mud, cryptocrystalline quartz with a cement of micrite and iron oxides, giving rise to darker or brownish laminae, while the grainstone laminae are calcite-cemented and light-grey in colour. The occurrence of thin grainstone and packstone laminae suggests that the detrital grains had been reworked by infrequent low-velocity currents and deposited in deeper water, possibly in the middle or outer shelf.

Trace fossils are frequently observed in these rocks and commonly occur on the upper surface of beds (i.e. exogenetic type), but endogenetic types are also present occurring within beds of argillaceous limestone. Most of the material of the burrows is recrystallised and replaced by micrite and/or sparry calcite and locally cryptocrystalline quartz.

Burrowing has greatly disrupted and modified the original internal structures of these sediments. In outcrops the burrows are randomly oriented and of various patterns with size grades up to 2 cm wide and up to tens of centimetres in length (Plate 2.5E, F). The presence of these trace fossils suggests a very low rate of sedimentation and/or periods of non-deposition during which the compaction of sediments could take place. A low rate of sedimentation is also suggested by the occurrence of hard ground features in some beds of the argillaceous limestone. The presence of iron oxides in all rock types might also suggests a low rate of sedimentation. Although the present distribution of the iron oxides is probably due to secondary diagenetic processes or weathering, their

Plate 2.6 Photomicrographs of the Sinsidik Beds.

A. Photomicrograph of glauconitic quartz-rich calcarenite from the Sinsidik Beds (83 TO 69.1), showing the fine grain and angular shape of terrigenous quartz and the rounded shape of the glauconite pellets (greenish or yellowish). Crossed polars, 40X.



B. Photomicrograph of argillaceous limestone of the Sinsidik Beds (83 TO 69.3), showing large amounts of terrigenous quartz silt, and calcite in a lime mud matrix. A few grains of glauconite are also present. Crossed polars, 40X.

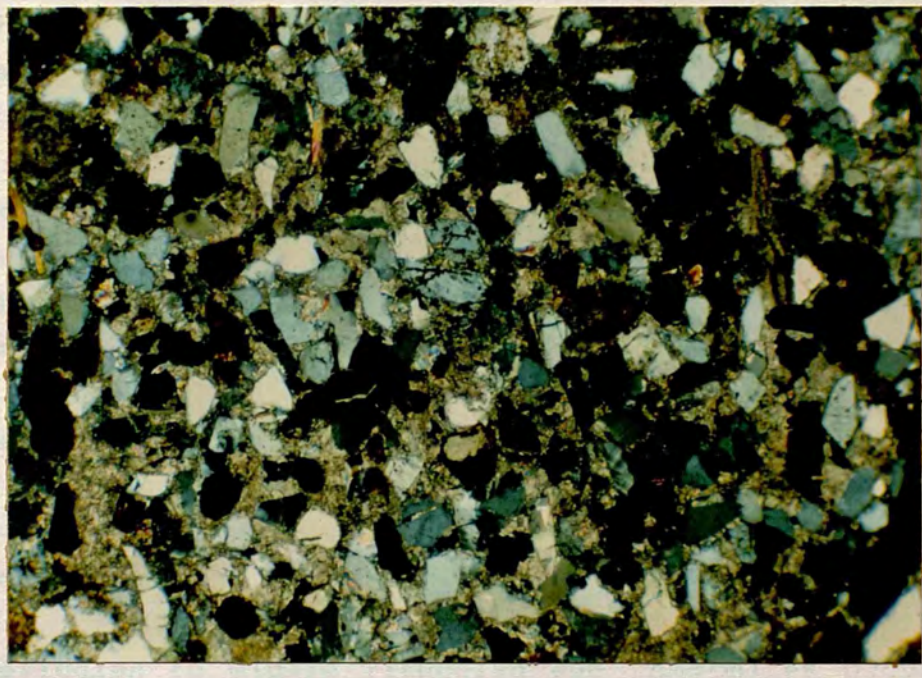


Plate 2.6C. Photomicrograph of calcareous lithic arenite of the Sinsidik Beds (83 TO 62.1) showing angular and monocrystalline terrigenous quartz and a few grains of prismatic muscovite. Note the micritised fossil fragments, probably molluscs (near top left corner and centre). Crossed polars, 40X.

occurrence may indicate that these rocks originally had a relatively high Fe-content, either as micronodules or in a dispersed form.

C. Stratigraphic relationship between subfacies

Although the succession is discontinuously exposed, due to the highly deformed and tectonised nature of the rocks, the outcrops suggest a vertical gradation between lithological types. The exposure along the coast near Luok shows that red limestone, marlstone and dark-grey calcarenite are interbedded, and in Tanjung Sinsidik these rocks are transitionally overlain by argillaceous limestone, which becomes abundant towards the top of the succession (Fig.2.5B). The lithological association also suggests a vertical gradation from carbonate containing terrigenous quartz in the lower part, to calcareous pelagic sediments in the upper portion of the succession.

D. Biostratigraphy

Belemnopsis uhligi Stevens (Plate 2.5H) occurring in limestone and marlstone was identified by Mr D. Phillips, Natural History, British Museum (personal comm.) as Kimmeridgian to middle Tithonian age. Ammonites occurring in marlstone and calcarenite were too crushed for generic identification, but their rib-pattern suggests several genera ranging from Oxfordian to Tithonian age (Dr. H.G. Owen, Natural History, British Museum, personal comm.).

The recrystallised calcareous foraminifera, some of which are identified by Bizon et al. (person. comm.) as ***Epistomina cf moquensis*** and ***Lenticulina sp.***, are of Callovian to Oxfordian age. The absence of siliceous foraminifera suggests that these microfaunas are pre-Cretaceous age.

The occurrence of belemnites, ammonites and calcareous planktonic foraminifera indicate a Late Jurassic age for the Sinsidik Beds, but the exact age range is still uncertain. The firmest age is assigned as Oxfordian to middle Tithonian.

In the Banggai-Sula islands, a similar but relatively much less-deformed argillaceous limestone-dominated sedimentary succession contains a belemnite-bivalve facies which includes Belemnopsis spp., Inoceramus spp., Buchia sp., and Malayamaorica malayamaorica(Krumbeck) of Late Oxfordian to middle Tithonian age (Sato et al., 1978; Westermann et al., 1978). This succession is included in the Buya Formation (Surono & Sukarna, 1985; Supanjono & Haryono, 1985).

E. Discussion and Interpretation

The depositional setting of the mixed red limestone and marlstone, grey calcarenite and light-grey argillaceous limestone is interpreted as a relatively deep continental shelf, probably middle to outer neritic depth, in which significant amounts of continentally derived quartz detritus was present. The presence of significant amounts of glauconite pellets and a few grains of apatite suggests a depth of deposition ranging from 30 to 1000 metres. This depth range is also suggested by the fauna of macroinvertebrates.

There is, however, some indication that the faunas, especially the belemnite guards were transported. Some of the belemnites were deposited at a low angle to the bedding planes, which suggests that the fragments were reworked and deposited by infrequent low-velocity currents. Angular terrigenous quartz, as large as fine sand in the calcarenite and marlstone, is considered to be derived and reworked from (inner) shelf deposits. The quartz grains have been transported downslope by infrequent low-energy currents which are also responsible for producing the laminae in calcarenite beds.

The occurrence of belemnites, ammonites and bivalves within a fine matrix of hemipelagic clays and the occurrence of abundant exogenetic trace fossils also suggests that these rocks were deposited in a deep shelf environment (Blatt et al., 1980).

The occurrence of argillaceous limestone in notably less amounts than the red limestone and marlstone may be attributed to the carbonate deposition being very slow, since the explosive proliferation of calcareous nannoplankton in Tithonian times had not yet occurred (Garrison and Fisher, 1964).

The argillaceous limestone containing numerous ghosts of micritised microfaunas was accumulated above the carbonate compensation depth (CCD). Van Andel (1975) estimated that the depth of Late Jurassic to Early Cretaceous CCD ranges from 3500 to 4500 metres for the open ocean. The CCD should be much shallower along the continental margins, because of high productivity of phytoplankton and high CO_2 levels in the pore waters of the resulting organic-rich sediments, which are not favourable to carbonate preservation (Leggett, 1985). If we accept that the red limestone and marlstone which contain belemnites, ammonites and bivalves and significant amounts of glauconite pellets had been deposited in the greater depth (i.e. middle to outer shelf), it is necessary to have a somewhat deeper sea or basin for deposition of the calcareous microfossil-bearing wackestone concordantly and subsequently on top of the red limestone and marlstone. This feature, further, implies that the basin was deepening toward the end of Jurassic times. This feature fits the change of sea-level proposed by Vail et al. (1977), which show a relatively rising of the eustatic sea-level towards the end of Jurassic time (Fig. 2.6.3). Tectonic implications of this palaeobasin will be further discussed in Chapter 5.

2.2.4 NAMBO BEDS

A. Definition

Nambo Beds is an informal name assigned to the highly deformed and imbricated marly-limestone dominated sequence exposed in Nambo river to the north of Nambo village (Fig.2.7). The unit is faulted against the Salodik Formation and its lower and upper portion are not seen.

B. Synonymy

The unit is named the Nambo Formation (Rusmana et al., 1984; Surono et al., 1984).

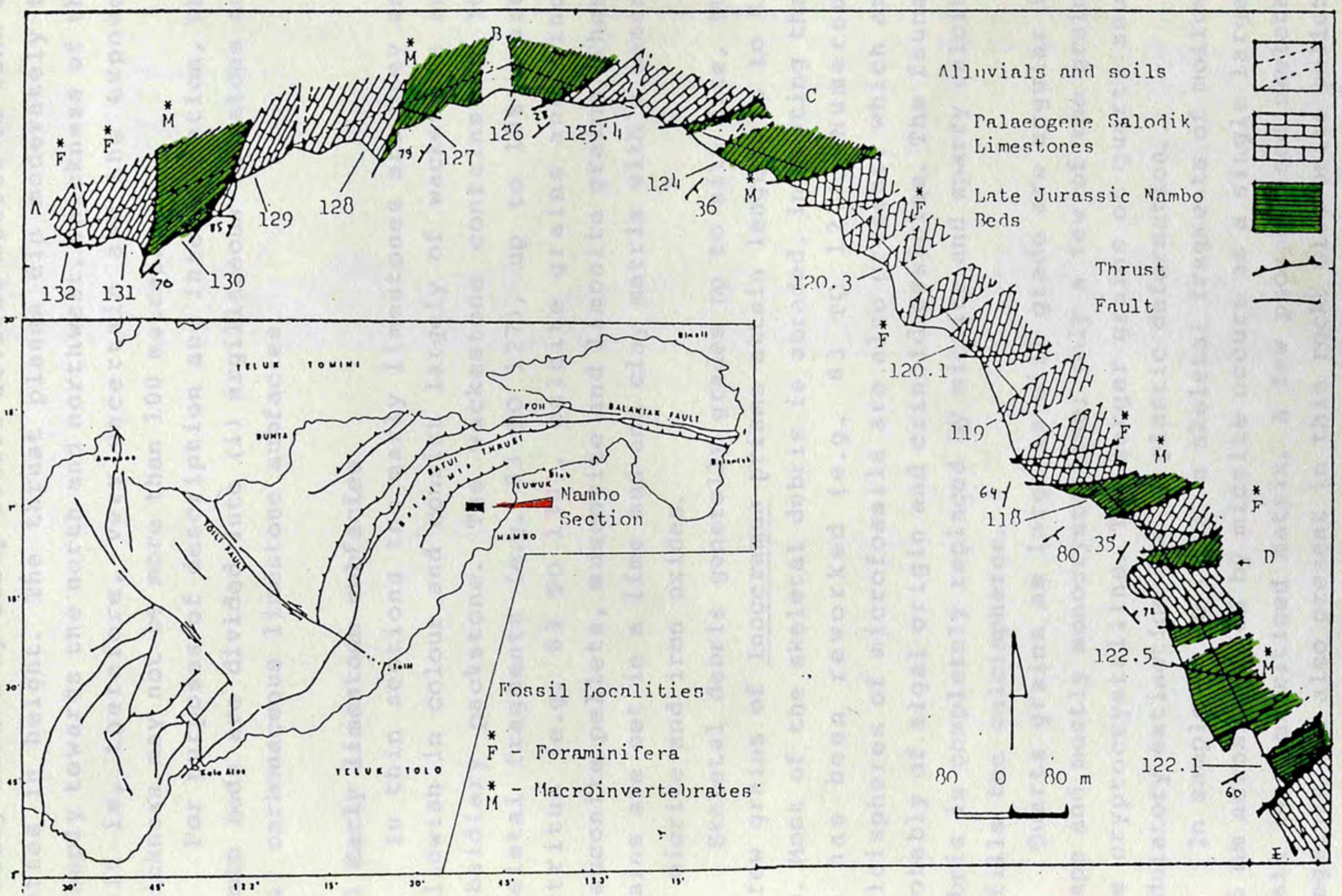
C. Description

The Nambo Beds occur in fault sliver exposures in Nambo river, to the north of the Nambo village. They are poorly exposed due to intensively weathering and crushed due to the imbricated nature of the rocks. The unit is dominated by fossiliferous, light grey to dark marly limestone. Bedding is poorly preserved, probably the bedding was disrupted by bioturbation, which is frequently observed occurring within these rocks. Beds up to 1 m thick with sharp base contacts are present and suggest that the rocks were originally thickly bedded or massive.

Macroinvertebrates including molluscs, bivalves and echinoderms occur in these rocks. Most of the fossils are intensely crushed and the belemnite guards are commonly laid down subparallel or at low angle to the bedding planes and show a weak imbrication, which suggests reworking by infrequent currents.

The outcrops are always in fault contact with the Salodik Limestones and slickensides showing vertical movement are commonly found within this unit. Thrusts are

Fig. 2.7 Geological traverse map of the Nambo River section.



usually marked by steep cliffs several metres to tens of metres in height. The thrust planes dip moderately to steeply towards the north and northwest. Thickness of the unit is, therefore, very uncertain and the exposed thickness may not be more than 100 metres.

For purposes of description and interpretation, the Nambo Beds are divided into (i) argillaceous limestone and (ii) carbonaceous limestone subfacies.

(i) Marly limestone subfacies

In thin sections the marly limestones are grey and yellowish in colour and consist largely of wackestone and subsidiary packstone. The wackestone contains up 30% skeletal fragments (e.g. 83 TO 127), up to 10% quartz detritus (e.g. 83 TO 130), calcite grains and minor glauconite pellets, muscovite and limonite grains. These grains are set in a lime mud and clay matrix with a cement of micrite and iron oxides.

Skeletal debris generally grades up to silt size, but a few grains of Inoceramus prisms attain lengths up to 0.3 mm. Most of the skeletal debris is abraded, indicating that it has been reworked (e.g. 83 TO 127). Numerous calcispheres of microfossils are also present, which are probably of algal origin and crinoid ossicles. The faunal debris is completely replaced by micrite and sparry calcite infills the calcispheres.

Quartz grains as large as silt grade are angular in shape and mostly monocrystalline; only a few of the grains are cryptocrystalline. The larger grains of quartz show undulatory extinction due to plastic deformation.

In sample 83 TO 125.3 a skeletal fragments of mollusc 1.5 mm across filled by micrite occurs as a single larger grain in a micritised matrix. A few phosphatised skeletal fragments are also present in this rock. Glauconite pellets of silt size may form up to 1% of the rocks (e.g. 83 TO 122.3). Some of the glauconite grains are rectangular

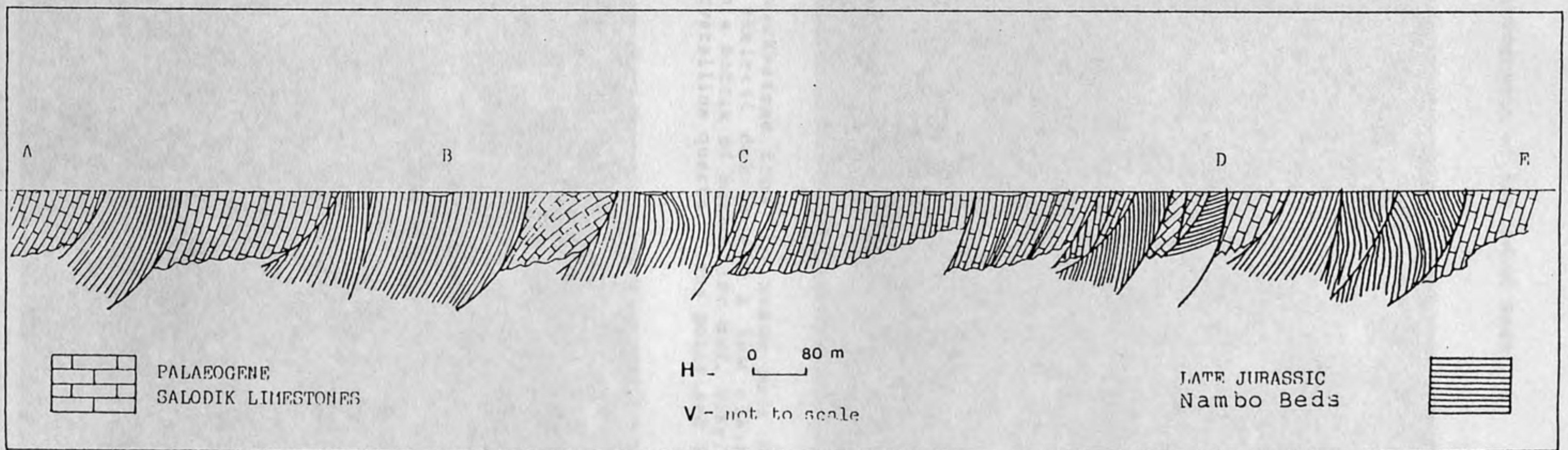
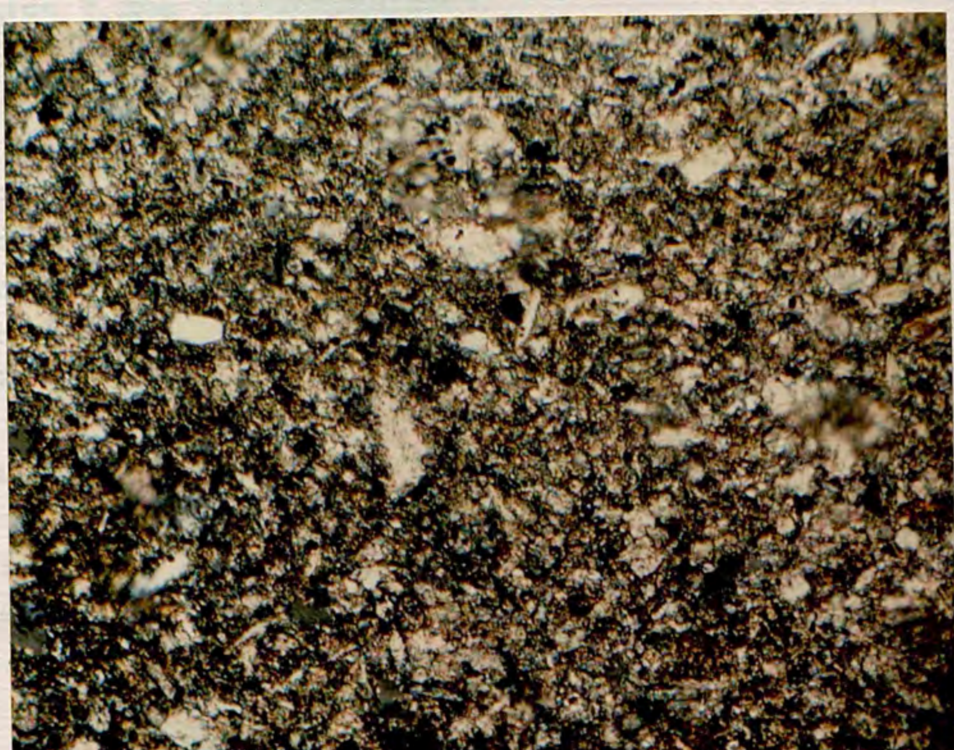


Fig. 2.8 Diagramatic line section of Nambo River section showing the imbricated nature of the successions.

Plate 2.7 Photomicrographs of the Nambo Beds

A. Photomicrograph of wackestone from the Nambo beds (83 TO 130.2), showing abraded skeletal debris and a few grains of terrigenous quartz set in a matrix of mixed lime mud, clay, iron oxides and minor cryptocrystalline quartz. Plane polarised light, 125X.



B. Photomicrograph of packstone from the Nambo beds (83 TO 127), showing the abraded nature of the skeletsl debris indicating reworking of the fragments. Note also the presence of a few grains of glauconite (greenish yellow). Plane polarised light, 125X.

shaped suggesting that they were formed from alteration of biotite. A few prismatic muscovite flakes may be present in some rocks.

The packstone contains up to 80% grains (e.g. 83 TO 128.2) consisting largely of skeletal fragments with very minor quartz detritus, set in a limonitic micritised lime mud matrix. The suite of grains is very similar to the wackestone, except for the much higher content of iron oxides and the smaller amounts of glauconite pellets.

Sedimentological features of these rocks, including the presence of a macroinvertebrates faunas, the presence of a significant amount of glauconite pellets and the occurrence of endogenetic bioturbation, strongly suggest that the rocks were accumulated in a shallow shelf depositional setting (Ahr, 1973; Tucker, 1985). However, the abraded nature of the skeletal fragments, the presence the significant amounts of angular quartz detritus, as well as few muscovite flakes and the weakly imbricated alignment of the belemnite guards all suggest that these rocks were transported and deposited in the deeper portion of the continental shelf. This will be discussed later in this section (see Interpretation).

(ii) <u>Carbonaceous limestone subfacies</u>

This facies consists of fine-grained, structureless, although sometimes thinly laminated, lime mudstone and highly recrystallised argillaceous limestones. They are characterised by a dark-grey to dark brown colour, absence of bioturbation, the presence of abundant carbonaceous matter and the presence of authigenic pyrite. These features all point to deposition under anoxic conditions.

A scattering of lime mudstone pebbles (chips) up to 10 cm long, subangular to angular in shape also occur in some beds. Bedding generally ranges in thickness from 5 to 25 cm, but a few beds up to 1 metre thick are also present.

The base and upper contacts of each bed are sharply defined. Macroinvertebrates including molluscs and echinoderms occur in the rocks; most of the fossils are intensely crushed. Some of the shells especially the bivalves, are lying subparallel or at low angle to bedding planes.

In thin section, the carbonaceous limestones are dark to brownish wackestone and mudstone, consisting largely of skeletal debris generally up to silt size, but a few larger grains of arcuate or elongate shaped up to 1.5 mm long (probably Inoceramus prisms) are present (e.g. 83 TO 118.2). Most of the skeletal fragments are micritised and a few grains are infilled by sparry calcite. All grains show abrasion features, indicating the reworking of skeletal debris.

Organic matter may be present, up to 5% and usually angular to subangular in shape and grades up to silt size. A few elongate grains, up to 0.3 mm long are also present in some rocks (Plate 2.7C). In sample 83 TO 122.5, most of the organic matter was pyritised. Berner (1970) on the basis of the sulphur isotope analysis, deduced that the sulphur in the pyrite of modern muds comes from two sources, i.e. (i) organic matter and (ii) sulphate dissolved in seawater. The organic matter is typically opaque and dark brown and reddish brown in colour, giving rise to the dark colour of the sediments.

Additional grains include the siliceous ghosts of microfossils and micritised calcispheres of probably algal and foraminiferal ogigin. All these grains are set in a matrix consisting of mixed lime mud, cryptocrystalline quartz and clays. Locally the matrix is quite siliceous.

D. Biostratigraphy

The Nambo Beds contain belemnites and bivalves. Most of the fossils were too crushed for generic identification.

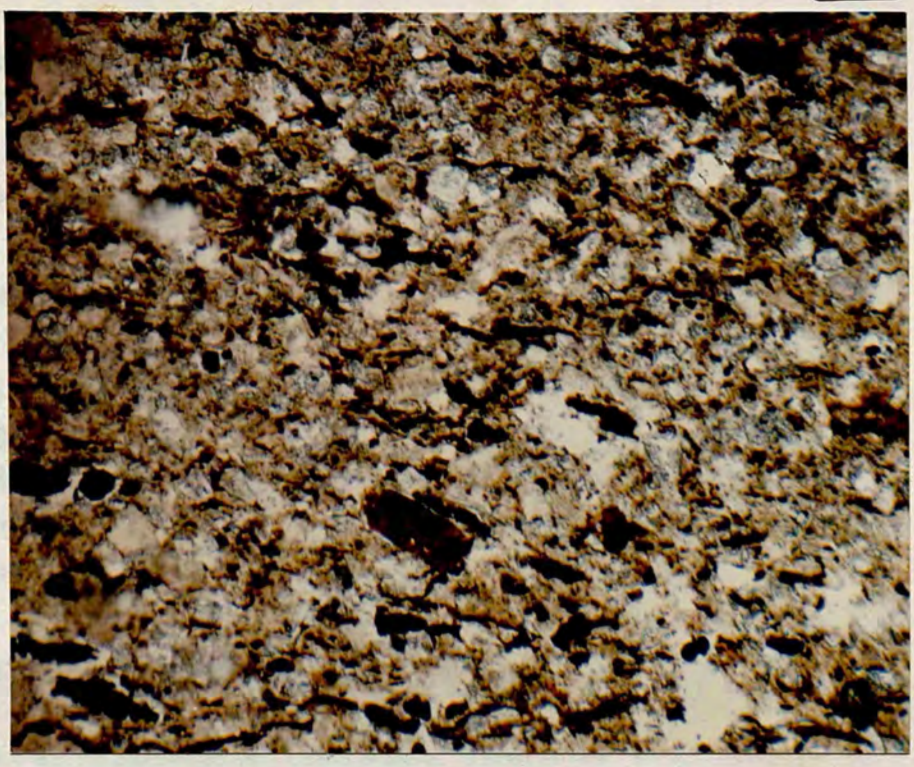


Plate 2.7C Photomicrograph of carbonaceous argillaceous limestone from the Nambo Beds (83 TO 118.2), showing the presence of large amounts of organic matter, giving rise to the dark colour of the rocks. Crossed polars,125X.

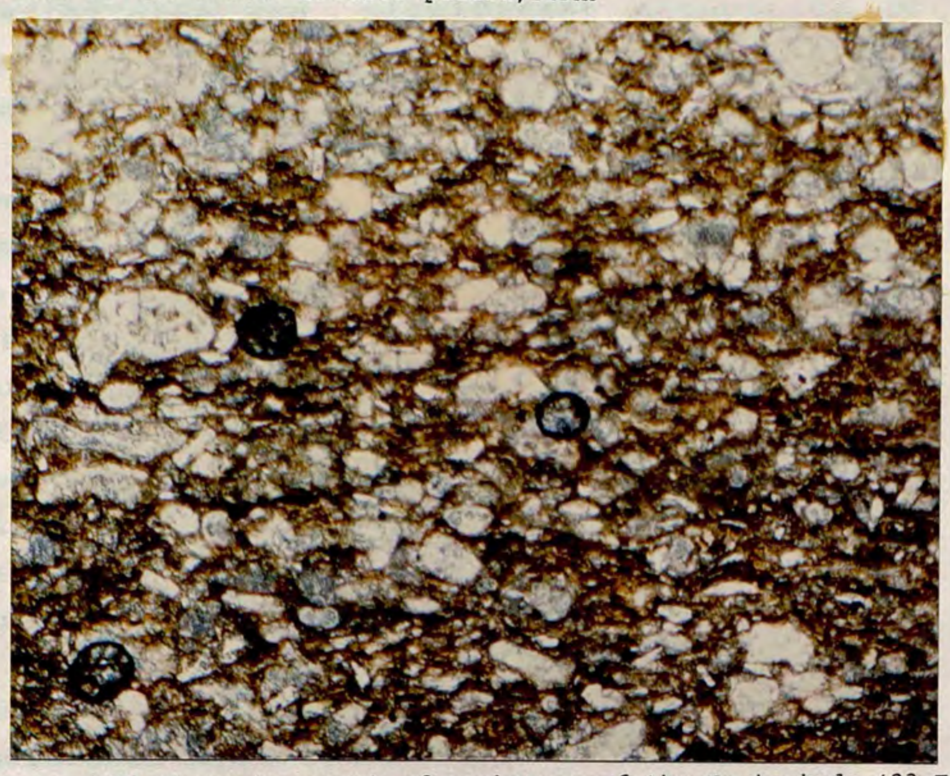


Plate 2.7D Photomicrograph of packstone of the Nambo beds (83 TO 123.2), showing the elongated fossils grains due to development of pressure solution cleavage. Some of the fossil grains show eyed-structure. The dark rounded spots are bubbles. Plane polarised light, 125X.

Mr D. Phillips and Dr.H.G. Owen, Natural History, British Museum (person. comm.) have identified Belemnopsis uhligi Stevens of Kimmeridgian to middle Tithonian age.

In the Banggai-Sula islands a similar succession consisting of shale and marls (i.e. Buya Formation, Surono and Sukarna, 1985), contains a belemnite-bivalve facies, which includes Inoceramus gracilis, Inoceramus haasti, Inoceramus stoliczkci, Malayomaorica malayomaorica, Haplophylloceras strigile and Blandfordiceras cf. wallichi of Kimmeridgian to Tithonian age (Sato et al., 1978; Westermann et al., 1978; Surono and Sukarna, 1985; Supanjono and Haryono, 1985).

E. Interpretation

The marly limestone contains abundant belemnites and subsidiary bivalves indicating a starved bank facies on a continental margin. The presence of abundant macroinvertebrates and significant proportion of glauconite pellets suggests that the depositional setting was at a depth between 30 to 1000 metres.

Some of the features of the marly limestone, including the presence of angular silt grade quartz detritus and abraded skeletal grains, coupled with the weak imbrication of the belemnite guards, strongly indicate that these fragments (grains) were reworked and deposited by infrequent low velocity bottom currents down slopes into deeper water in the middle or outer shelf.

The presence of angular pebbles of calcareous mudstone in the base of some beds clearly indicates the reworking of the sediments. The pebbles are considered to be derived from inner shelf deposits.

The occurrence of siliceous-rich carbonaceous wackestones intercalated with marly limestones suggests a rapid subsidience of carbonate shelf below the carbonate compensation depth (CCD) or alternatively and more likely, the limestone has been redeposited and interbedded with the siliceous wackestone which accumulated below the CCD in the deeper shelf environment.

Apart from the diagenetic features above, the major differences between the marly limestone and the carbonaceous argillaceous limestone subfacies are the dark and reddish brown colour and scarcity of bioturbation in the carbonaceous argillaceous limestones, indicative of deposition under very low oxygen conditions. These features suggest that during deposition and development of the thinly laminated lime mudstone and wackestone, the bottom waters were completely anaerobic and a burrowing infauna was excluded. Under reducing conditions, the sediments were enriched in organic matter, resulting in dark colour of the rocks. The high organic content led to the reduction of sulphates to sulphides in the form of pyrite, during later diagenesis. Jenkyns (1980) has suggested that the Callovian-Oxfordian was period of an 'Oceanic Anoxic Event', because organic-carbon-rich marine deposits of that age occur throughout Tethyan sequences, as well as in other parts of Europe (Hallam and Bradshaw, 1979).

Laminated marlstone beds of Tethyan affinities generally contain more than 5% organic matter and are considered to have been deposited in a deep oxygenated environment. Some of the lamination may represent reworking by bottom currents (Bernoulli and Jenkyns, 1974; Weissert et al., 1979; Arthur and Premoli Silva, 1982). The marly limestones of the Nambo Beds are similar to those Tethyan sediments in both lithological and sedimentological features, and therefore, were considered to have been deposited in relatively deep shelf, probably in the middle to outer shelf.

2.2.5 LUOK BEDS

A. Definition

Radiolarian calcilutite with chert nodules occurring in the highly deformed and imbricated fault-slivers along the coast near Luok village is informally named the Luok Beds (Fig.2.9). The upper and lower boundaries of the unit are not seen in the type locality.

B. Description

The Luok Beds occur in fault-bounded exposures along the coast near Luok village, some 4 km to the north of the Balantak, and as blocky exposures in the hilly topography along the coast of Tanjung Padingkian to the north of the Luok village (Fig. 2.9). The unit is faulted against the Sinsidik Beds.

The Luok Beds are dominated by a succession of pink, white and pale-grey calcilutite with red or grey chert and/or cherty calcilutite nodules. They are even-bedded with bed thickness ranging from 4 to 25 cm, dense, hard and compact or lithified. The upper and lower contacts of beds are sharply defined. Thin parallel laminae, usually less than 2 cm thick, occur in some beds of calcilutite and commonly curve around the chert nodules. Microfossils occur in most of the rocks.

The unit is highly deformed; bedding changes abruptly in both attitude and dip over very short distances. Near Luok village, the rocks are steeply dipping with a fairly well developed spaced pressure solution cleavage, which disrupts the laminae (Plate 2.7D). In Tanjung Padingkian, bedding suddenly changes from nearly vertical to horizontal reflecting a possible monocline fold developed during thrust movement (Plate 2.7C). Joints and fractures are the most significant mesoscopic structure occurring in these

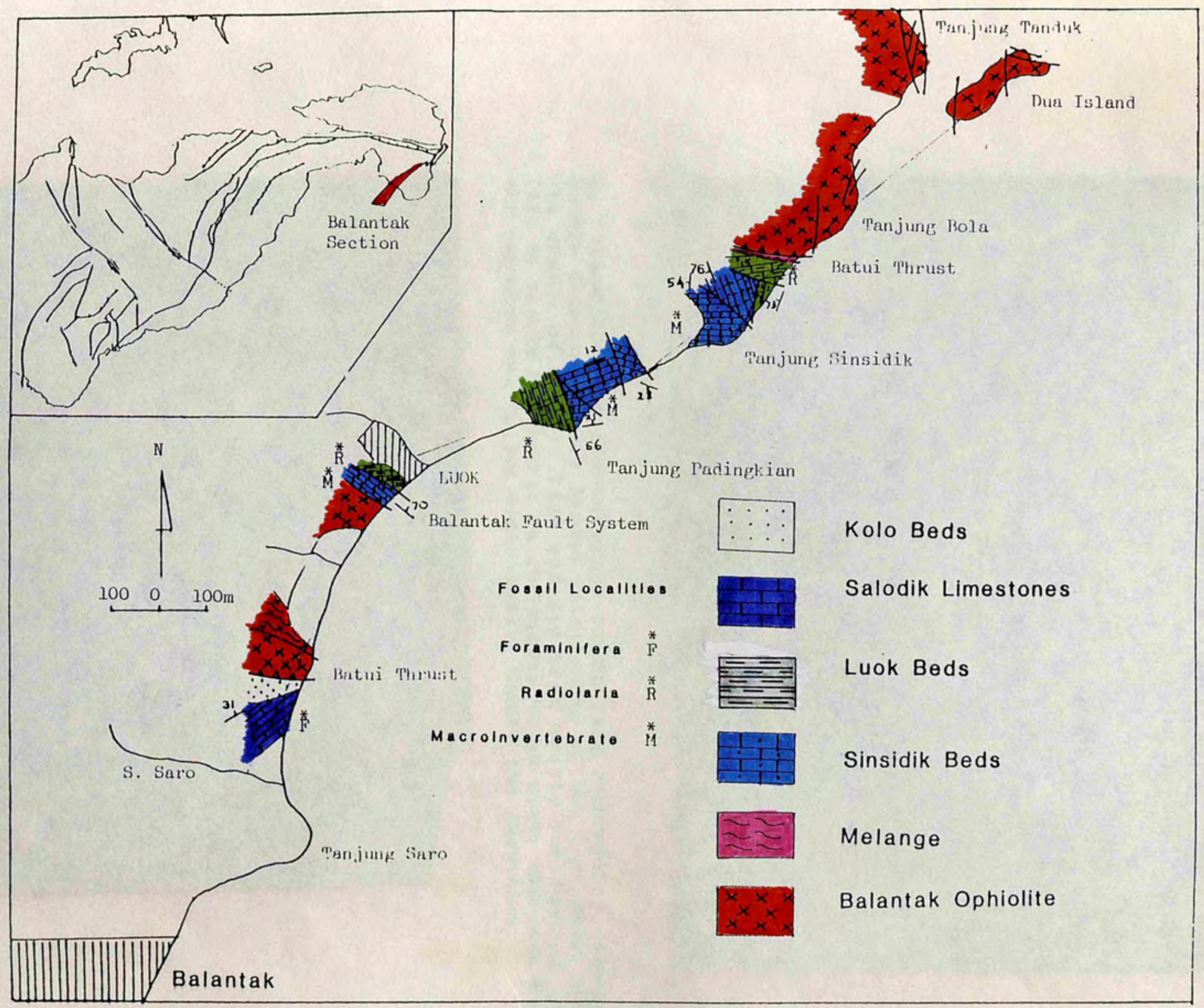


Fig. 2.9 Geological traverse map of Balantak coast.

Plate 2.8 Photographs of the exposures of the Luok Beds

A. Photographs of the outcrops in Tanjung Padingkian showing highly deformed thinly bedded calcilutite of the Luok Beds. Note the change of bed attitude and dip from nearly vertical on the left to nearly horizontal on the right, which might indicate that the sequence was folded monoclinally during the thrusting.

B. Photograph of the calcilutite of the Luok Beds exposed near Luok village, showing development of pressure solution cleavage. The cleavage is penetrative in both handspecimen and under microscope.

rocks. The outcrops suggest that these structures have occurred repeatedly indicating a multiplicity of deformation. Thickness of the unit is very uncertain, but the exposed thickness is not more than 50 metres. Contact with the underlying and overlying units are not seen.

Calcilutite

In outcrops the calcilutite is highly deformed and intensely fractured, showing smooth and conchoidal fracture surfaces with a porcellaneous lustre. In thin sections the calcilutite is pale grey and yellowish wackestone and packstone and consists largely of calcareous microfossils and minor grains of silt size skeletal debris including molluscs, echinoids and a few grains of crinoid ossicles and ostracods (e.g. 83 TO 62.1). Most of the microfossil tests are walled-calcispheres, which are largely filled by micrite and in some cases by sparry calcite. In some rocks, calcispheres may make up to 50% of the rocks (e.g. 83 TO 65.1). These grains are set in a matrix of a mixed lime mud, clays and minor cryptocrystalline quartz, cemented by micrite and iron oxides. Sparry calcite is also locally present as cement.

Thin parallel laminae of several millimetres up to 2 cm thick, are defined by the alternation of calcispheres-dominated packstone and wackestone or lime mudstone laminae. The calcisphere-packstone is cemented by micrite and iron oxides resulting in a dark-brownish colour, while the wackestone is calcite-cemented and light-grey in colour (e.g. 83 TO 65.2). In some rocks, the laminae have been disrupted by bioturbation and also by pressure solution cleavage, which have transformed the original laminae (e.g. 83 TO 62.2). The cleavages cut through the calcispheres of microfossil. Later fractures cut the cleavage perpendicularly and are filled by sparry calcite.

The molluscan and echinoderm fragments may form up to

C

D

Plate 2.8C&D Photographs of the outcrops in Tanjung Padingkian showing the occurrence of the chert nodules within the calcilutite sequence of the Luok Beds. The nodules may be over 1 m long (D), and are mostly occur in oval shape (C).

20% of the rocks, usually show abraded features, indicating that they have been reworked by low velocity bottom currents (Burchette & Britton, 1985).

Chert Nodules

In outcrops the chert nodules occur in various colours of light-grey, brownish-grey and reddish brown and are intensely fractured (Plate 2.8C). The nodules occur in various sizes and shapes, ranging from a few centimetres in egg-shaped nodules, to 30 cm in lensoidal nodules, often flattened in the plane of the bedding. In some cases, the nodules may occur up to over 1 metre long (Plate 2.8D). Nodule exteriors are commonly bleached, soft and porous due to alteration at some stage in the history of the nodules. In general, nodules are concentrated along particular bedding planes and are nearly absent along adjacent ones. Locally, chert nodules are so abundant that they coalesce in the plane of bedding, forming discrete beds with uneven surfaces, which are up to 3 m thick and several metres in length. The larger nodules usually show a knobbly exterior.

In thin sections the chert nodules are grey and reddish in colour, and consist largely of microcrystalline quartz and numerous ghosts of microfossils, cemented by amorphous silica with extremely low birefringence and iron oxides. Most of the microfossil calcispheres are replaced or filled by dark-brownish Fe-Mn oxides giving rise to the reddish colour of the cherts. A few crystalline siliceous radiolarian tests are also present (e.g. 83 TO 65.2), and indicate that the chert nodules have been formed from crystallisation of chemically unstable amorphous silica. Some of the nodules are opaline, containing numerous microscopic specks of carbonate, which are abundant at the contact zone with the calcilutite host rocks. This feature indicates that the host rocks were incompletely replaced by silica (Plate 2.8E). The chert nodules were formed very

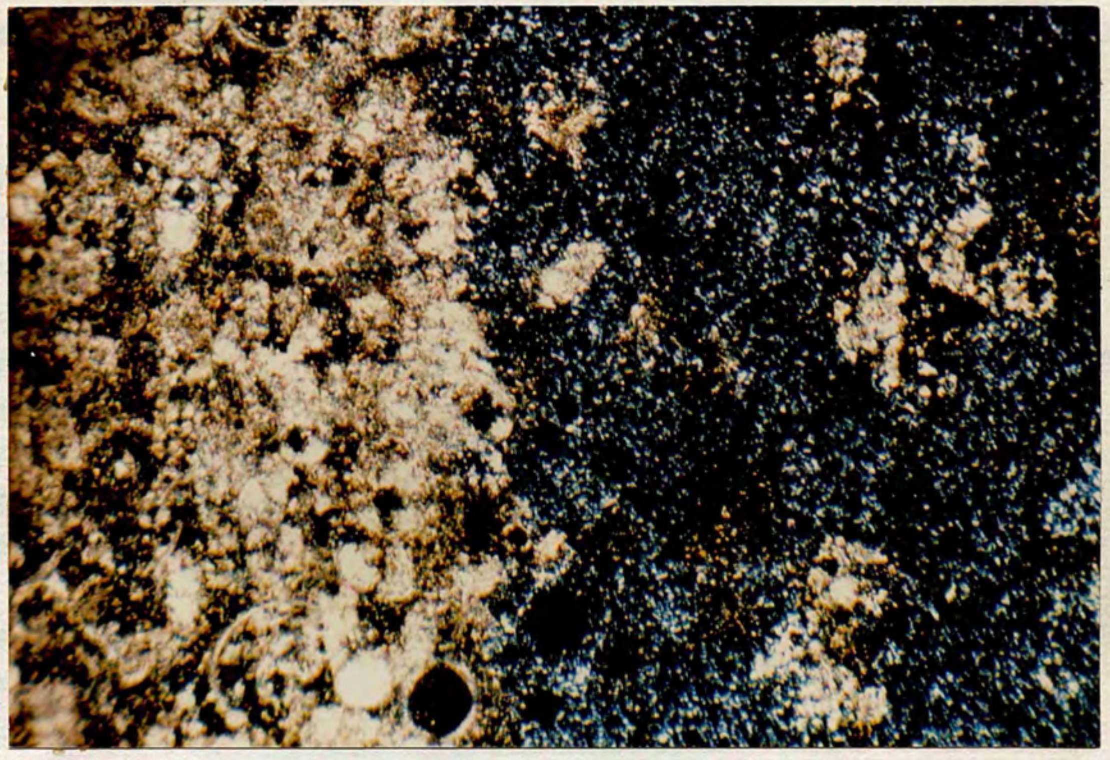


Plate 2.8E Photomicrograph of chert nodule in the Luok Beds, showing numerous microscopic specks of carbonate at the contact zone with the calcilutite host rocks. This feature indicates that the carbonate host rocks were incompletely replaced by silica.

early during diagenesis of the calcilutite, as the laminae are curved around the nodules, showing that compaction of the host rock postdated nodule formation.

C. Biostratigraphy

The calcilutite of the Luok Beds is characterised by presence of abundantly but variably preserved calcareous nannoplankton and planktonic foraminifera. The nannoplankton faunas were identified by Bizon et al. (person. comm., 1983), include Arkhangelskiella cymbiformis, Eiffelithus turriseiffeli, Cribrosphaerelle ehrenbergi, Prediscosphaera cretacea, Micula staurophora, Microrhabdulus decoratus, Tetralithus aculeus and Watznaueria barnesea of Campanian to Late Maastrichtian age.

The planktonic foraminifera were identified by Purnamaningsih-Siregar of GRDC (person. comm.) and Bizon et al. (person. comm., 1983), and include Globotruncana sp., Globotruncana cf. falso-stuarti and Heterohelix sp. of Turonian to Maastrichtian age. Kundig (1957) reported the occurrence of Cretaceous foraminifera in argillaceous limestone on the Biak-Poh road, about 10 km to the north of Salodik village, which in this study is included within the Luok Beds. The foraminifera include Arenobulimina sp., Globotruncana nosetta (Casey), Guembelina globulosa (Ehrenberg), Pseudotextularia frusticosa (Egger). Based on the microfaunas described above, the age of the Luok Beds, is therefore Campanian to Maastrichtian.

In the Banggai-Sula islands, a similar succession (i.e. the Tanamu Formation of Surono & Sukarna, 1985) consisting of alternating marly limestone and calcilutite with shale intercalations contains planktonic foraminifera, including Globotruncana sp., Globotruncana stepani-turbinata, Globotruncana cf. lapparenti, Globotruncana coronata and Globotruncana cf. fornicata of Cretaceous age.

Nannoplankton are also present, among them, Arkhangelskiella cymbiformis, Broinsonia parca, Eiffelithus turriseiffeli, Tetralithus corpulatus, Microshabdulus decoratus, Watznaueria barnesea and Lucianorhabilus cayeuxi of Cretaceous age (Surono & Sukarna, 1985; Supanjono & Haryono, 1985).

D. Interpretation

The sequence of carbonate rocks in the Luok Beds is typically pelagic, consisting of calcareous nannoplankton and microfossils (lithified calcareous ooze). The most significant sedimentary features of these rocks include thin and parallel-sided beds, sometimes with parallel laminae, and the occurrence of chert nodules. They rarely show textural or compositional evidence for reworking and/or redeposition of the sediments. These are entirely of fine grain pelagic sediments. Terrigenous and carbonate detritus derived from shelf deposits are nearly completely absent.

The lithological association and sedimentological features of these rocks strongly indicate that the Luok Beds were accumulated at bathyal depths, above the carbonate compensation depths (CCD), on the distal portion of a passive continental margin. Based on back-tracking DSDP site data of Berger and Winterer (1974), Van Andel (1975) showed that the CCD was in the range of 3200 to 4000 metres for the Pacific and Indian Oceans during the Cretaceous to the Palaeogene. The CCD, however, is much shallower along the continental margin due to the high productivity of the phytoplankton and the high CO_2 levels in the pore waters of the resulting organic-rich sediments which are not favourable to carbonate preservation (Leggett, 1985). It is, therefore, reasonable to estimate that the Luok Beds were accumulated in a sea or basin with a maximum possible depth of 3000 to 3500 metres.

Several features of the sedimentary record of the Luok Beds may be related to regional palaeogeographic and tectonic factors. The most conspicuous is the Early Cretaceous hiatus. This hiatus is also recorded in the Banggai-Sula Islands (Surono & Sukarna, 1985; Supanjono & Haryono, 1985), Misool Island (Rusmana et al., 1982; Pigram et al., 1983), Obi and Bacan Islands (Aswan et al., 1979; Pigram and Panggabean, 1983) and Kepala Burung, Irian Jaya (Pigram & Panggabean, 1983). The hiatus is an uncommon stratigraphic feature in the Early Cretaceous sedimentary succession of Tethyan affinities. Jenkyns (1977) summarised the Cretaceous sedimentary succession in many parts of the Tethyan Basin, which is characteristically dominated by pelagic rocks deposited in bathyal depths. Pigram & Panggabean (1983) have however also documented the occurrence of continuous sedimentation during Cretaceus time in some parts of eastern Indonesia, e.g. Birds Neck, East Irian Jaya and Papua New Guinea.

Vail et al. (1977) show a relative falling of global sea-level in Early Cretaceous time (Fig. 2.6.3). They suggest the eustatic sea-level in the Early Cretaceous fell abruptly for a short period after the end of Jurassic, but rose again toward the middle of Cretaceous. The depositional setting of the Luok Beds, therefore, fits the global change of sea-level during Late Cretaceous time.

2.2.6 SALODIK LIMESTONES

Definition

A carbonate-dominated sedimentary succession, deformed and imbricated, occurring in the southern portion of the East Arm of Sulawesi, in named the Salodik Limestones (Fig.2.9). The unit unconformably overlies Mesozoic sediments (i.e. Louk, Nambo and Sinsidik Beds) and is unconformably overlain by the Neogene coarse clastic sediments. It is everywhere faulted against ophiolite.

Synonymy and Derivation

The succession was named the Salodik and Poh Formations by Rusmana et al. (1984). The name is derived from Salodik village, where the unit is best exposed.

Description

The Salodik Limestones occupy most of the southern portion of the East Arm of Sulawesi (Fig.2.9) and also occur in Peleng Island. The rocks typically form a karst topography of hilly to mountainous country extending from Batui in the west to Balantak in the east.

The unit also occurs as isolated exposures in the Kolo Atas region. Subsurface data shows that the Peleng Strait at least partly is occupied by this succession, underlain by basal clastics and granitoid and metamorphic basement and overlain by coarse clastic sediments.

Stratigraphic relationship of Salodik Limestones with the Mesozoic sedimentary successions is not clearly seen in the East Arm of Sulawesi. The bisostratigraphy of the unit, however, suggests that the Salodik Limestones are unconformably underlain by the Mesozoic sediments (i.e. Luok, Nambo, Sinsidik and Kapali Beds). In most cases, the unit is in fault contact with the Mesozoic sediments and the ophiolite as well.

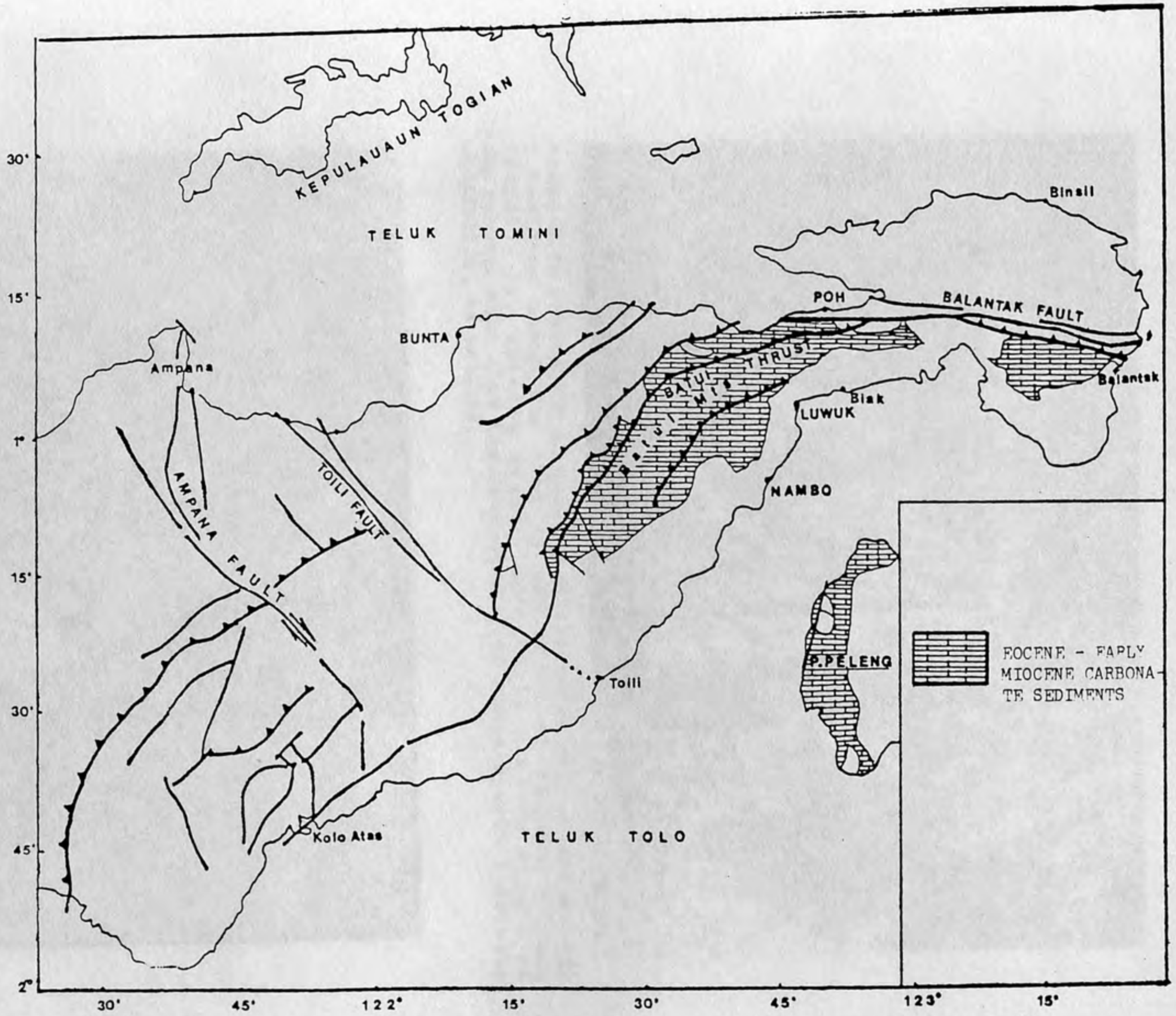


Fig. 2.10 Map showing structural configuration of the East Arm of Sulawesi and the occurrence of the Palaeogene Salodik Limestones.

Plate 2.9A Photograph of outcrop of the upper part of the Salodik Limestones along the road to the north of Salodik village (83TO111.1), showing nearly vertically dipping strata of marlstone. Note also the development of close-spaced fractures subparallel to bedding.

Plate 2.9B Close-up of photo A showing the marlstone beds with slickensides which indicate horizontal movement (red arrow showing sense of movement). The occurrence of this feature is related to the dextral movement of the Balantak Fault System, which occur not far away to the right of photo A.

In the Balantak area, the unit is faulted against ophiolites and Luok Beds. In the Biak-Poh section, the outcrops show that the unit is unconformably overlain by coarse clastic rocks (i.e. the Biak Conglomerates) and Quaternary coral reefs. In the Kolo Atas area, the unit occurs as blocky exposures and is in fault contact with the ophiolites and the Early Jurassic Kapali Beds. The outcrops suggest that the Salodik Limestones unconformably underlie the quartz-rich clastic sediments (i.e. Kolo Beds) and are tectonically bounded by melange (Kolokolo Melange).

In the Nambo River section, the unit is everywhere faulted against the Late Jurassic Nambo Beds and the fault contact is usually marked by a steep cliff several metres to tens of metres high. The rocks are intensely faulted and thrust; slickensides showing vertical movement are frequently observed in these rocks. In Peleng Island the unit unconformably overlies the granitoid and metamorphic basement complex (Surono et al., 1984). Aerial photographs show that the succession is intensely imbricated, with thrust planes moderately to steeply dipping to the north or northwest (see Chapter 5 Structures).

The Salodik Limestones are a carbonate-dominated sequence of a massive and thickly bedded, grey and yellowish-grey foraminiferal limestones with grey marlstone intercalations in the lower part of the formation. Towards the upper part, marlstone becomes gradually more abundant and is much better bedded with thickness ranging from 5 cm to nearly 1 metre. The marlstones show parallel-sided beds with a sharp basal contact. Parallel laminae occur in some beds of marlstone.

In outcrops the rocks are moderately to nearly vertically dipping to the north or northwest. The limestones in the lower portion of the unit are poorly bedded, commonly massive or thickly bedded, up to 6 metres thick. Most of the limestone beds are structureless, and the basal and upper contacts of each bed are sharply

Fig. 2.11 Geological traverse map of the Biak-Poh section.

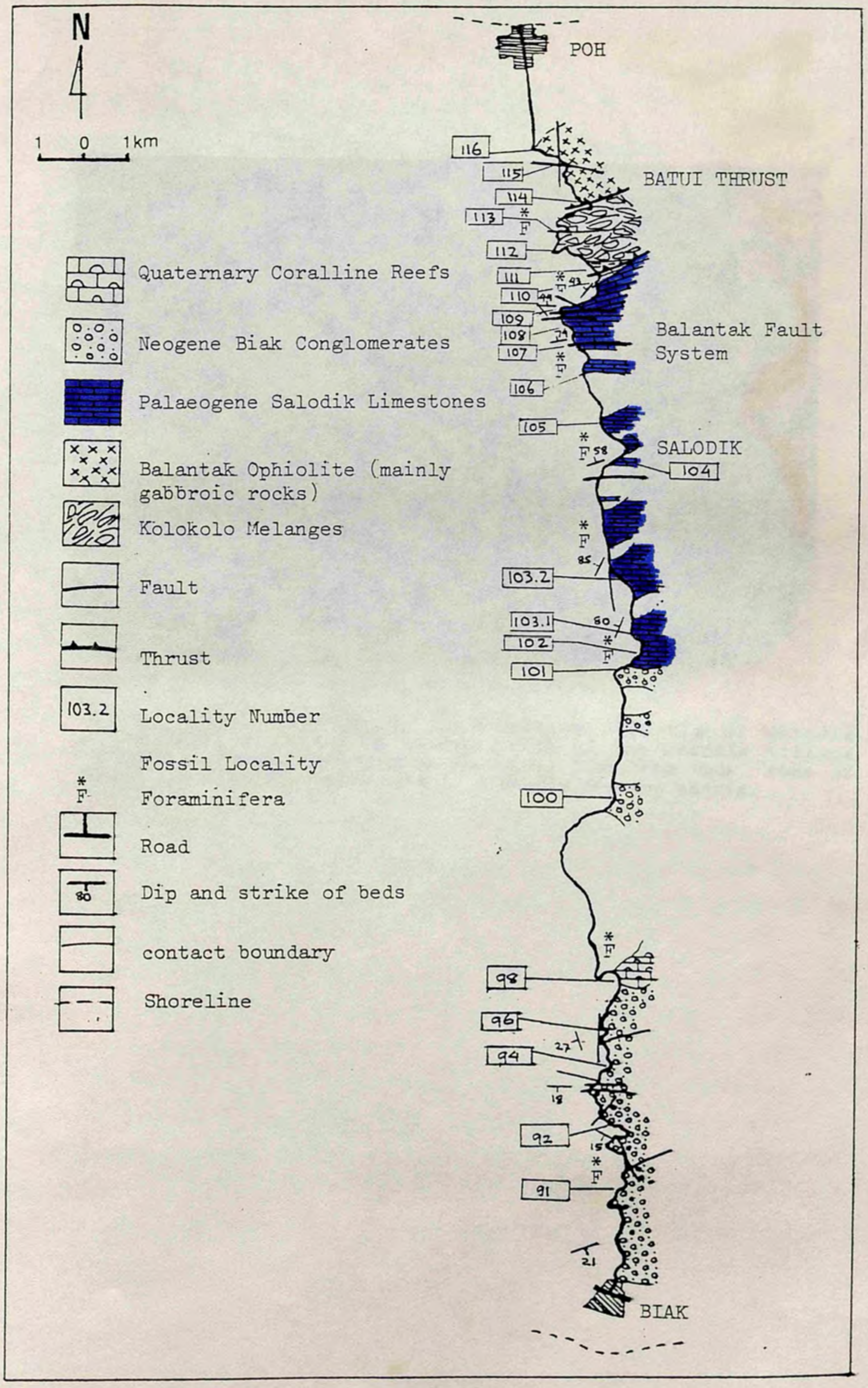


Plate 2.9C Photograph of road-cutting exposure of Salodik Limestones, about 12 km to the north of the Salodik village (83TO110), showing the highly fractured limestone beds, some of which are slightly boudinaged in a weakly sheared matrix.

defined.

Trace fossils occur in some beds of limestone, mostly of an endogenetic type, but with some exogenetic type preserved trails occurring on surface of beds.

Macrofossils including echinoids, molluscs and corals are present in some beds of limestone in he lower part of the succession. Larger benthic foraminifera occur in most of limestone beds and planktonic foraminifera occurs in most of marlstone and in some limestone beds (see Biostratigraphy).

In thin sections the carbonate sequence consists of grain-supported grainstone and packstone and matrix-supported wackestone. Boundstone limestone occurs locally in the Nambo river and Balantak sections.

Boundstone

The boundstone limestones are composed primarily of skeletal grains of macroinvertebrates and microfossils. In thin sections, the rocks are typically very dark, due to the presence of iron oxides in the matrix and as cement. They contain grains dominated by skeletal fragments of macroinvertebrates (e.g. 83 TO 128.2), including moluscs, algae and echinoids and subsidiary larger benthic foraminifera and calcispheres of planktonic foraminifera. In 83 TO 17, the skeletal grains consist largely of algae and echinoids and subsidiary benthic and planktonic foraminifera.

These skeletal grains are set in minor amounts of matrix consisting of fine grained calcite, lime mud and iron oxides, and are cenented by limonitised micrite. Although most of the skeletal grains are replaced by micrite or infilled by sparry calcite, the internal structures of some fossils, especially benthic foraminifera are still visible (e.g. 83 TO 64.1; 83 TO 128.2). Some rocks have a sparry calcite cement stained by iron oxides (83 TO 103).

Plate 2.10A Photomicrograph of foraminiferal limestones of Salodik Limestones (83 TO 107.2), showing benthic foraminifera. Plane polarised light, 40X.

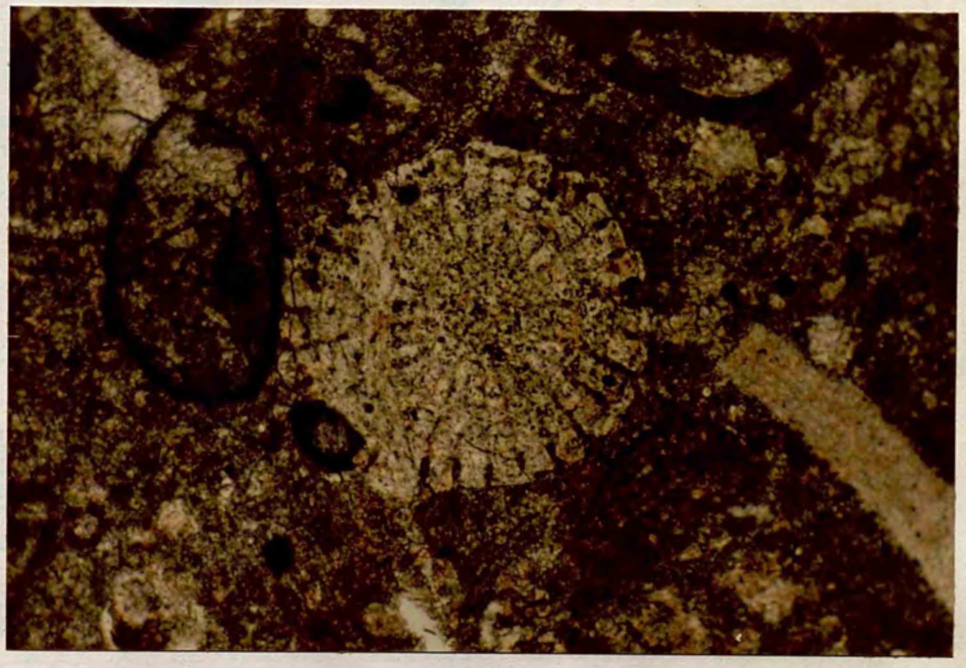


Plate 2.10B Photomicrograph of packstone of Salodik Limestones (83 TO 107), showing presence of echinoid spine (centre) in addition to the foraminifera. Plane polarised light, 40X.

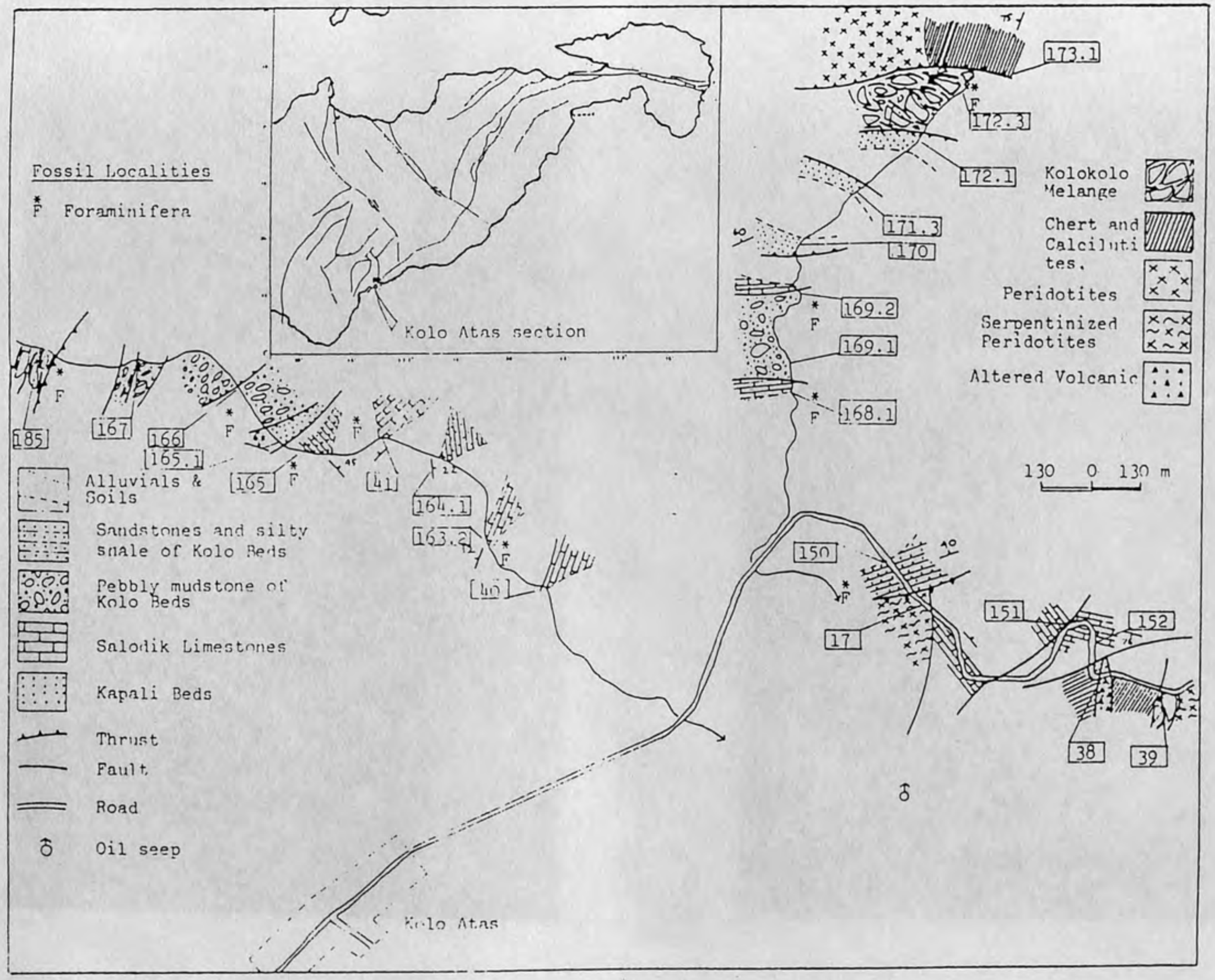


Fig. 2.12 Geological traverse map of Kolo Atas area,

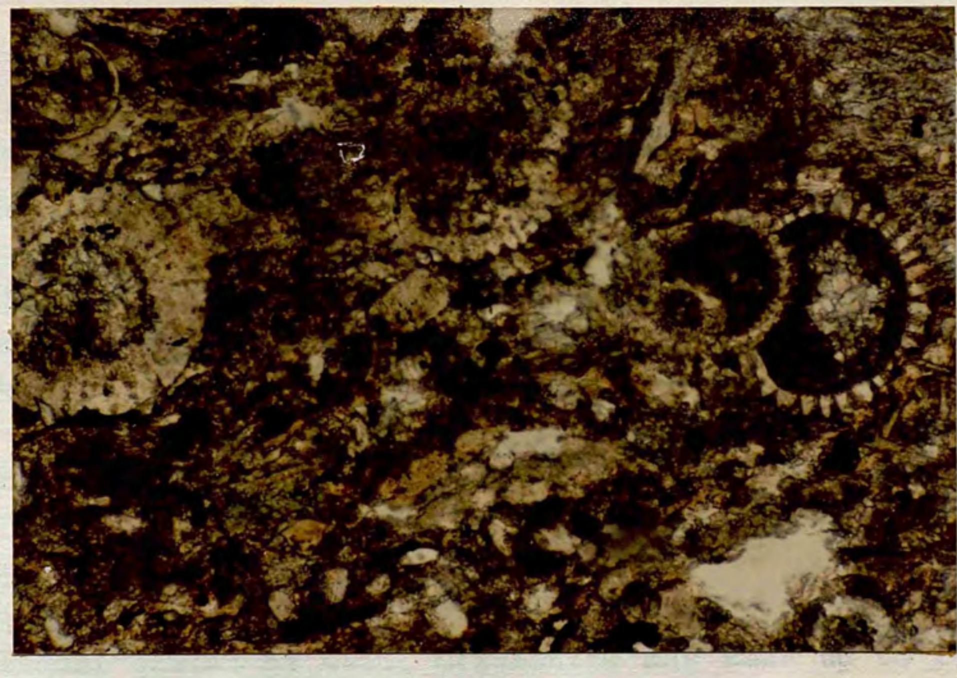


Plate 2.10C Photomicrograph of foraminiferal limestone of Salodik Limestones occurring in Tokala Atas area (83 TO 50B), showing presence of planktonic foraminifera, including Globigerinoides sp. (G), Sphaeroidinellopsis sp. (S) and Orbulina universa (O). Plane polarised light, 40X.

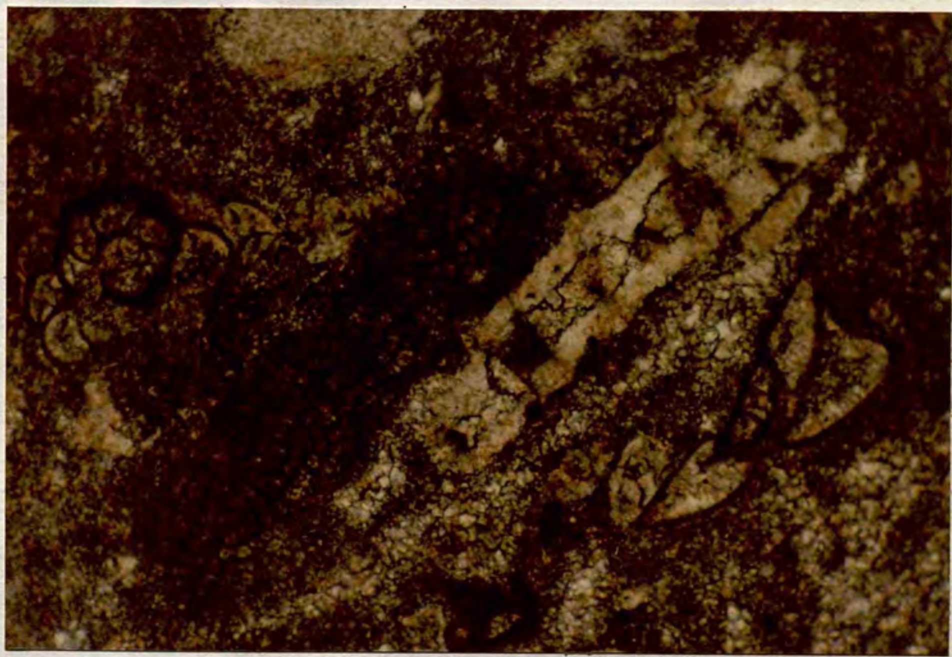


Plate 2.10D Photomicrograph of foraminiferal grainstone of Salodik Limestones occurring in Nambo River (83 TO 119.3), showing benthic foraminifera, i.e. Lepidocyclina sp. (L), Operculina sp. (O) and coralline algae as well (C). Plane polarised light, 40X.

Grainstone Limestones

The grainstones are typically grain-supported limestone and contain skeletal debris making up to 70% of the rock (e.g. 83TO 107.2), consisting predominantly of benthic foraminifera and algal fragments, echinoids and calcispheres of planktonic foraminifera. Skeletal debris of molluscs is also present. Most of the skeletal grains are replaced by micrite and infilled by sparry calcite. The internal structures of the fossils, particularly the larger benthic foraminifera and algae, are still recognisable.

Additional grains, present in small amounts, include monocrystalline quartz of silt size and glauconite pellets also of silt size (e.g. 83 TO 30; 83 TO 35.2). In slide 83 TO 35.2 the glauconite pellets occur inside some of the benthic foraminifera (i.e. Nummulites) and as separate individual grains as well. Glauconite may make up to 2% of the rock. These grains are set in a matrix consisting of lime mud and micrite with a cement of carbonate and iron oxides. A sparry calcite cement is also present in some rocks (e.g. 83 TO 62.3).

Packstone Limestones

The packstone limestones compositionally are identical to the grainstones, the differences are that texturally they contain more lime mud matrix more quartz detritus and glauconite pellets, but a smaller amount of macroinvertebrate skeletal fragments (e.g. 83 TO 17; 83 TO 34.1; 83 TO 94; 83 TO 101A; 83 TO 103; 83 TO 107.2; 83 TO 111.1). The skeletal fragments consist largely of larger benthic foraminifera, with subsidiary planktonic foraminifera, algae, echinoidea and minor skeletal debris of unidentified macroinvertebrates. The skeletal fragments are mostly replaced by micrite and some are infilled by sparry calcite. Internal structures of benthic foraminifera are visible in some slides, and some of the planktonic foraminifera are micritised with interiors infilled by lime mud (e.g. 83 TO 111.2).

The terrigenous quartz is mostly monocrystalline, and may constitute up to 10% of the rock (e.g. 83 TO 111.1). Cryptocrystalline quartz is also present as matrix or cement (e.g. 83 TO 109.1). Glauconite pellets are present up to 5% (e.g. 83 TO 34.1), with size grades up to 0.3 mm across; some of them are phosphatised, characterised by a yellowish-brown colour and are typically isotropic (Plate 2.11A). Some of the benthic foraminifera cells are filled by quartz detritus mixed with lime mud, which suggests that the terrigenous quartz was reworked and redeposited during and subsequent to the accumulation of the faunas.

__Wackestone Limestone__

In outcrops the wackestone occurs as well-bedded marlstone which gradually increases in proportion toward the upper part of the succession. In thin sections, the wackestone are grey or yellowish in colour and contain up to 30% skeletal debris (e.g. 83 TO 117.1; 83 TO 119.2), consisting largely of both planktonic and benthic foraminifera and subsidiary algae, echinoid and minor molluscan debris (Plate 2.10 F). Most of the skeletal grains are replaced by micrite or infilled by sparry calcite. Quartz detritus of silt size is present and may constitute up to 10% of the rocks (e.g. 83 TO 119.4). The quartz consists mostly of monocrystalline grains and is subangular to angular in shape. The larger grains show undulatory extinction and strained features due to plastic deformation. Minor cryptocrystalline quartz may be present locally as a matrix. The rocks are well cemented by carbonate and iron oxides.

The thin laminae are defined, microscopically by the alternation of skeletal-dominated packstone and matrix-supported wackestone. The packstone laminae are usually much darker in colour owing to the presence of much larger amounts of iron oxides as grain euhedra and as cement.

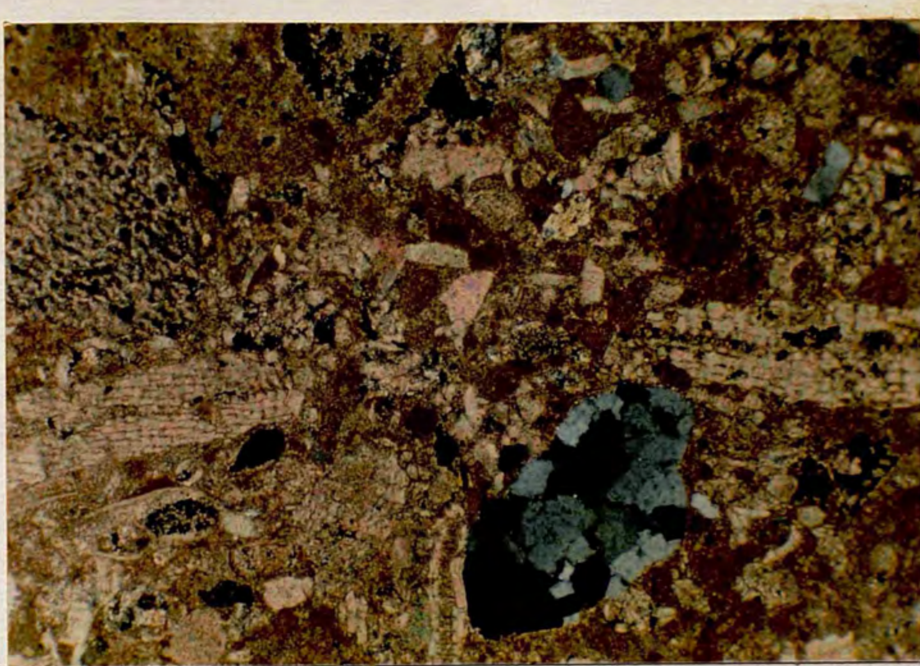


Plate 2.10E Photomicrograph of coralline limestone from the Balantak coast (83 TO 35.2), showing a few grain of terrigenous quartz, some of which show undulatory extinction due to plastic deformation. Crossed polars, 40X.

PLate 2.10F Photomicrograph of grainstone of Salodik Limestones occurring near Salodik village (83 TO 101A), showing planktonic foraminifera, i.e. Globigerinoides sp. and a few grains of terrigenous quartz. Plane polarised light, 40X.

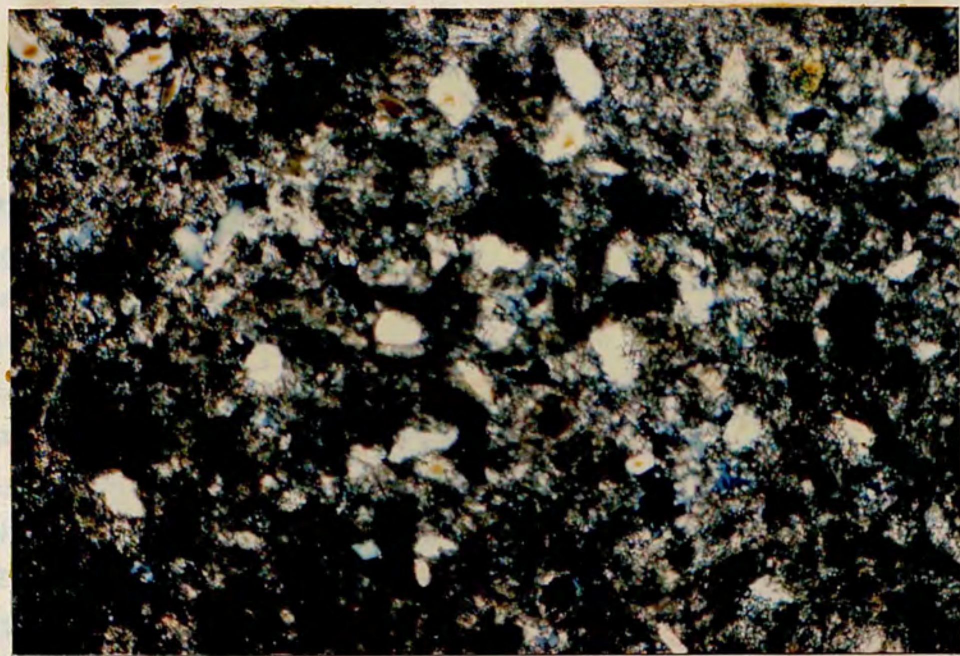


Plate 2.11A Photomicrograph of marlstone of Salodik Limestones (83 TO 111), showing significant amounts of terrigenous quartz, which are typically angular to subrounded in shape. Note also the presence of planktonic foraminifera and other skeletal debris. Plane polarised light, 125X.

Plate 2.11B Photomicrograph of marlstone of Salodik Limestones (83 TO 109.1), showing a few grains of glauconite (greenish and yellowish spots). Note also large amounts of subrounded to angular grains of quartz. Plane polarised light, 125X.

D. Biostratigraphy

The Salodik Limestones contain an abundance of both benthic and planktonic foraminifera. Rock samples collected from over 50 localities were studied and the microfossils identified by the Paleontological Laboratory of the Geological Research and Development Centre (GRDC), Bandung, Indonesia. Several rock samples collected from Biak-Poh and Balantak sections were examined by Mrs G. Bizon and her colleagues.

Age determination is based on the foraminiferal biostratigraphic zonations of Blow (1969), and Postuma (1971).

In the Biak-Poh section (e.g. 83 TO 103) the grainstone and packstone limestones contain benthic foraminifera which include Lepidocyclina sp., Operculina sp., Miogypsina sp. and Miogypsinoides sp. of Early to Middle Miocene age. In the locality 83 TO 104, the packstone limestone contains benthic foraminifera including Lepidocyclina sp., Cycloclypeus sp., Operculina sp., Miogypsina sp. and Miogypsinoides sp. of Early to Middle Miocene age. In the locality 83 TO 107.2, the packstone limestone contain both benthic foraminifera, including Nummulites sp., Spiroclypeous vermicularis and planktonic foraminifera, including Globorotalia sp. and Globigerina sp. (Plate 2.10.A). This faunal assemblage indicates a Late Eocene age for these rocks. The packstone of 83 TO 109.2 contain abundant planktonic foraminifera, including Globorotalia aragonensis, Globorotalia spinuloinflata, Globorotalia centralis, Truncorotaloides rohri and Truncorotaloides topilensis of Middle Eocene age.

In the Balantak section, the packstone and grainstone limestones (83 TO 30; 83 TO 35.2; 83 TO 35.6) contain benthic foraminifera, including Nummulites sp., Heterohelicidae sp., Spiroclypeus cf vermicularis and planktonic foraminifera which include Globorotalia cf.

spinulosa, Globigerina sp., Chilogumbelina, Globigerinatheka and Globigerinidae. Nannoplankton are present in the wackestone and they include Discoaster saipanensis, Cyclicargolithus floridanus, Sphenolithus meriformis, Dyctyococcites dictyodus, Coccolithus copelagicus and Micrantholithus vesper. This faunal assemblage indicates a Middle to Late Eocene age for these rocks.

In the Nambo river, the packstone and grainstone limestones (e.g. 83 TO 118.1; TO 119.1; 83 TO 120; 83 TO 129; 83 TO 132) contain abundant benthic foraminifera, including Lepidocyclina sp., Spiroclypeus sp., Operculina sp., Sphaeroidinellopsis sp., Rotalia sp., Lepidocyclina parwa, Cellanthus coraticulatus, Miogypsina spp., Ammonia beccavini and planktonic foraminifera which include Globigerina sp. and Globorotalia sp.. This faunal assemblage gives a Late Oligocene to Early Miocene age for these rocks.

In the Kolo Atas area, the packstone and grainstone limestones (83 TO 30; 83 TO 36A; 83 TO 150; 83 TO 168.3) contain benthic foraminifera, which include Operculina sp., Cycloclypeus sp., Lepidocyclina sp. and planktonic foraminifera, including Orbulina universa, Globorotalia menardii, Globigerinoides obliquus, Globorotalia cultrata, Globigerinoides immaturus Bronnimann, Globoquadria altispira (Cushman & Jarvis). This foraminiferal assemblage indicates an Oligocene to Early Miocene age for these rocks.

Mr. C.P. Nuttall (Natural History, British Museum, person. comm., 1985) examined the macroinvertebrates collected from the Nambo River and Kolo Atas area. The fossils consist mostly of molluscs which are intensely crushed and have lost their shells. Determinations are therefore mainly at family or generic level. Many of the gastropods are canaliculate which shows that they belong to either the more advanced Mesosgastropoda or the

Plate 2.11C Photomicrograph of Heterostegina sp. occurring in grainstone limestone of the Salodik Limestones (e.g. 83TC30).

Plate 2.11D Photomicrograph of Lepidocyclina sp. occurring in boundstone limestone of the Salodik Limestones (e.g. 83TO64.1).

Neogastropoda. Several genera are present, including Terebellum, Strombus, Cyparea and Conus and the family Cassidae which are common in the Neogene and in recent faunas, but also occur in older rocks. Vulsella, a genus of bivalve, is common in the Eocene sediments of the Indo-Pacific region.

Based on these faunal assemblages, particularly the benthic and planktonic foraminifera, the age of the Salodik Limestones is Early Eocene to early Middle Miocene (P7 to N12-13).

E. Stratigraphic relationship between lithofacies

The boundstone and grainstone limestone beds are concordantly overlain by packstone which are intercalated with wackestone beds in the Nambo river section. In the Balantak section, however, the boundstones and grainstones occur in fault slivers and blocky exposures, hence their stratigraphic position is not clearly known.

In the Biak-Poh section, the boundstone limestone is not exposed, but the grainstones conformably underlie the packstone limestone beds. Towards the upper part of the formation, the packstones are interbedded with wackestone limestones, and they also occur as alternating packstone and wackestone laminae in some beds of marlstone. These outcrops suggest that the facies pattern of the Salodik Limestones consists of a gradational change from boundstone-grainstones to packstone-wackestones from the lower to upper portion of the succession.

F. Discussion and interpretation

The facies pattern of a carbonate platform commonly consists of tidal flat deposits comprising skeletal-peloidal wackestones with lenticular grainstone in the platform interior, skeletal-peloidal grainstones and packstones of the shallow subtidal (above fairweather wave-base) and skeletal packstones and wackestones with grainstone horizons in the deeper subtidal (below fairweather wave-base). Below storm wave-base, skeletal wackestone dominates with thin beds of storm-derived packstone-grainstone limestones (Tucker, 1985).

The Salodik Limestones generally fit this facies association, especially in the lower part of the succession. The sedimentological features and lithological association of the Salodik Limestones, however, strongly suggest that there was a gradual and consistent change of facies in both vertical and lateral directions.

Compositionally and texturally, the Salodik Limestones show a gradual change from the grainstone-boundstone dominated limestones in the lower portion to the packstone-wackestone dominated limestones in the middle and upper parts of the succession, respectively. This is also signified by a gradual decrease in size of grains from the base to the upper part of the succession. These features suggest that the shelves were deepening, probably due to subsidence.

The deposition of the boundstone-grainstone facies probably occurred in the subtidal areas during storms; skeletal debris was transported, sorted and deposited to develop grainstone beds, while the winnowed shell lags would be left after the passage of storm currents to form the boundstone beds.

Sedimentologically, the formation shows that bedding gradually changed from poorly bedded and massive limestone in the lower portion, to much clearer parallel-sided beds with the development of parallel laminae in some marlstone beds, in the upper part of the succession. Palaeontologically, the unit shows that there is a gradual change from benthic foraminifera and benthic macroinvertebrates in places in the lower part to more abundant planktonic foraminifera in the upper portion of the succession. These features strongly indicate subsidence of the shelf.

Geographically, or perhaps it can also be said tectonically, the gradual change of these features can be traced from Peleng Island in the south as far as the northern part of the Poh Neck in the East Arm of Sulawesi. This present physiographical configuration does not appear to have changed much since Plio-Pleistocene time.

These features strongly suggest that, (i) the platform undoubtedly was part of an open shelf in Eocene times; (ii) the platform appears to have started subsiding during or after deposition of the lower part of the formation, giving rise to a deepening of the continental shelf. The planktonic foraminifera-rich marlstone and limestones of the upper part of the succession were deposited on a deepening shelf, which continued to subside through Oligocene to Middle Miocene times. This deepening shelf depositional setting is also suggested by the presence of significant amounts of terrigenous quartz in the marlstone beds, which indicates that the limestone were reworked from inner shelf deposits and transported and deposited downslope by infrequent bottom currents, which are also responsible for producing the laminae in the marlstone beds. Some of the skeletal debris of the macroinvertebrates shows abrasion features, indicating reworking of the fragments. The presence of glauconite pellets in the packstone and wackestone limestones suggests that the shelf from which these pellets were derived was in the range of 30 to 1000 metres depth in Oligocene-Middle Miocene times, during the deposition of the upper portion of the Salodik Limestones.

Morris (1979) described the occurrence and accumulation of carbonate sediments within a deepening basin in the Arabian Peninsula and southwestern Iran shelves. Deepening of this basin is due to collision of the African and Eurasian plates. There, the basin deepening started in late Early Cretaceous time (105 my) and continued until Miocene time (25 my) with a rate of sedimentation of 1 mm/40 y.

In Sulawesi, the outcrops of the Salodik Limestones suggest that the maximum thickness of the unit is about 1400 m. If the rate of sedimentation above is taken into account, the thickness of Salodik Limestones, therefore, is possibly in the range 1000 to 1200 metres. The present estimated thickness of the formation is much higher. The thickening of the unit is primarily due to the deformation and imbrication of the succession.

CHAPTER 3

STRUCTURES AND TECTONIC EMPLACEMENT OF THE BALANTAK OPHIOLITE

3.1 INTRODUCTION

This chapter contains a study of the Balantak Ophiolite, which represents the northern portion of the Eastern Sulawesi Ophiolite Belt (ESOB). The study includes structures, petrology and age determination of the ophiolitic rocks. The pelagic sediments (Boba Beds) which occur associated with the Balantak Ophiolite are also described in this chapter.

The Kolokolo Melange which occurs along the Batui Thrust-Balantak Fault System (i.e. surface expression of the collision zone between the BSP and ESOB), provides direct evidence of time and space, kinematics and tectonic emplacement of the ophiolite belt. The melange, therefore, will be described and discussed at the beginning of this chapter.

3.2 KOLOKOLO MELANGE

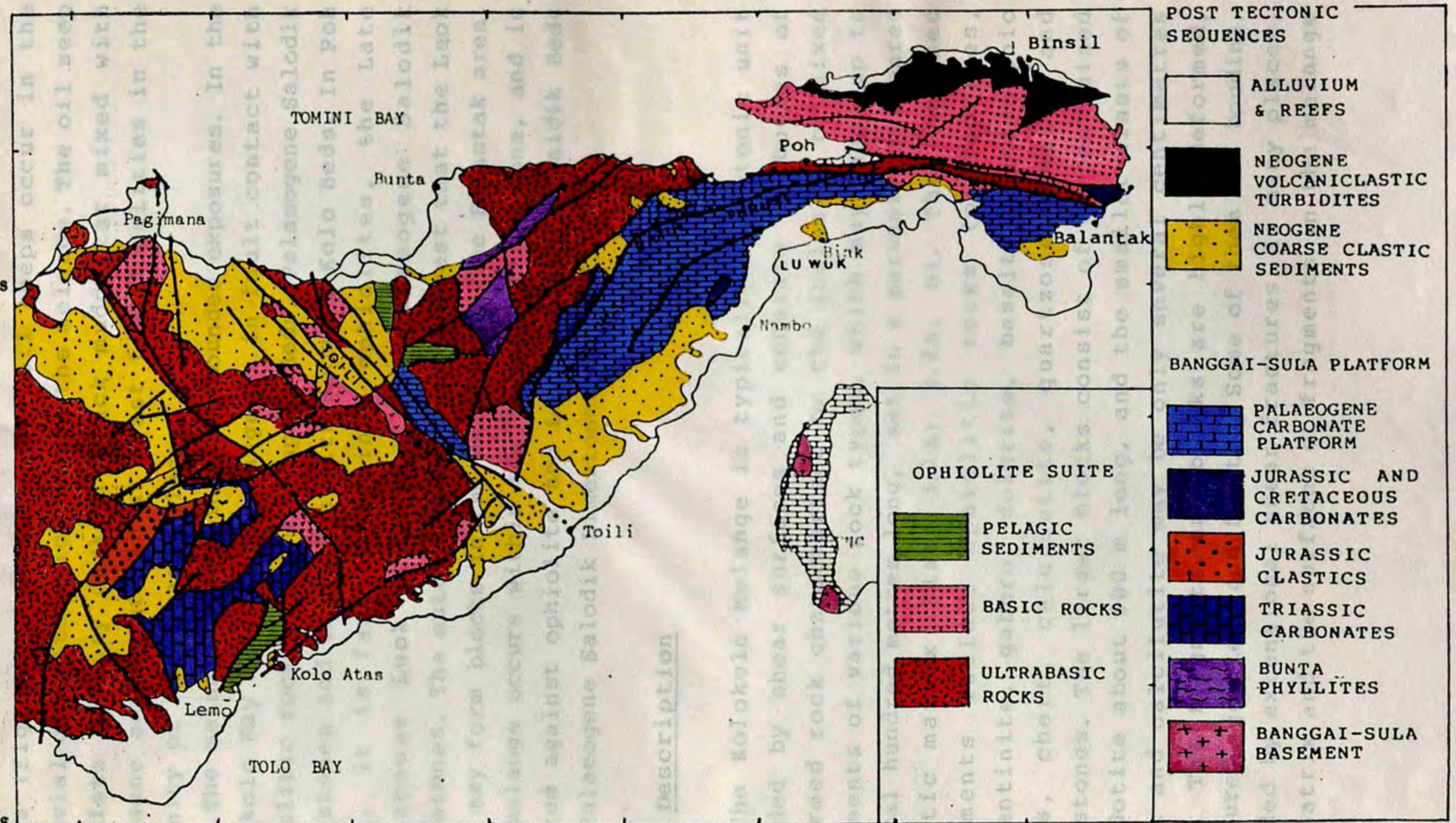
3.2.1 Definition

An assemblage of deformed rocks characterised by the inclusion of mixed fragments and blocks in a pervasively sheared matrix occurring in the vicinity of Kolokolo Bay is informally named the Kolokolo Melange. The unit is tectonically bounded by the adjacent rock units.

3.2.2 Distribution

The Kolokolo Melange crops out in the hilly topography surrounding Kolokolo Bay and as small exposures in many places along the Batui Thrust and Balantak Fault System In

Fig. 3.1 SIMPLIFIED GEOLOGIC MAP OF THE EAST ARM OF SULAWESI (After Simandjuntak et al., 1983; Surono et al., 1984; Rusmana et al., 1984)



the Kolo Atas area, the melange occurs associated with oil seeps (Fig. 3.2). In fact, the oil seeps occur in the alluvial deposits which covers the melange. The oil seep consists of very dark or black muddy tar mixed with kerosene and occurs in at least five localities in the vicinity of Kolo Atas village.

The melange occurs in fault bounded exposures. In the Kolokolo Bay area, the melange is in fault contact with ophiolitic rocks, Jurassic Kapali Beds, Palaeogene Salodik Limestones and in places with Neogene Kolo Beds. In Poh Neck, it is faulted against ophiolites, the Late Cretaceous Luok Beds and the Palaeogene Salodik Limestones. The outcrops, however, suggest that the Luok Beds may form blocks in the melange. In the Balantak area, the melange occurs within the Balantak fault zone, and is faulted against ophiolite and Late Jurassic Sinsidik Beds and Palaeogene Salodik Limestones as well.

3.2.3 Description

The Kolokolo Melange is typically a tectonic unit bounded by shear surfaces and consists of bodies of deformed rock characterised by the inclusion of mixed fragments of various rock types, which may range up to several hundred metres long, set in a pervasively sheared pelitic matrix (Plate 3.1A; 3.2A, B). The enclosed fragments include ophiolitic rocks (peridotites, serpentinite, gabbro, dolerite, basalt), metamorphic rocks, chert, calcilutite, quartzose arenite and limestones. The larger blocks consist of serpentinised peridotite about 500 m long, and the smaller clasts of chert and calcilutite may be only several centimetres long. The fragments and blocks are highly deformed, fractured, jointed and faulted. Some of them are boudins, bounded by extensional shear fractures. In many places, the matrix and the surfaces of fragments in the melange

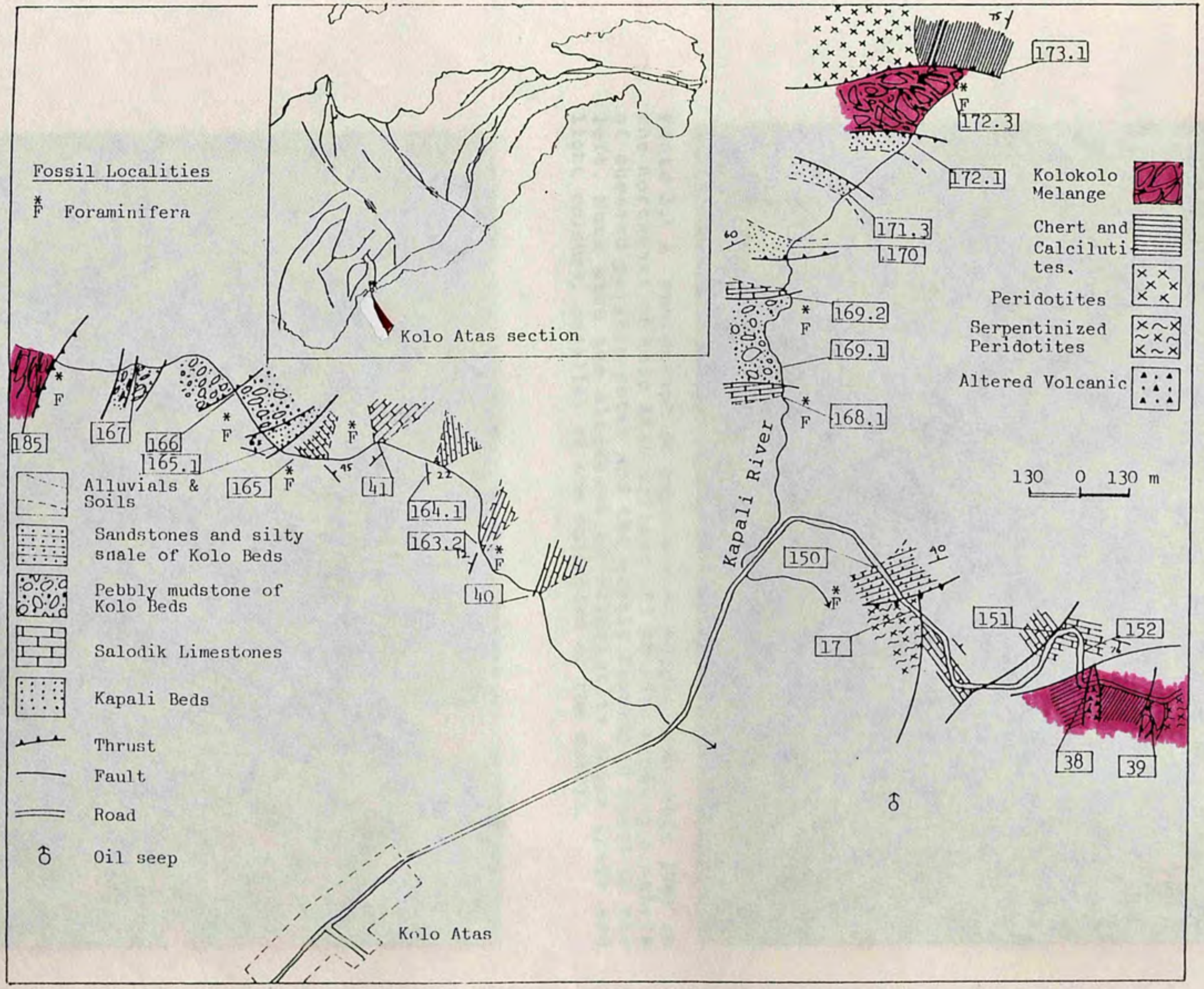


Fig. 3.2 Geological traverse map of Kolo Atas area showing the occurrence of Kolokolo Melange.

3.2

Plate 3.1 A Photograph of exposure of Kolokolo Melange just to the northeast of Kolo Atas village (83 TO 17), showing a matrix of sheared pelitic rocks and the highly fractured chert on the left. Note also the alignment of calcilutite chips (pale and light colour), parallel to the foliation of the matrix.

Plate 3.1 B Photograph of outcrop of Kolokolo Melange up-stream in the Kapali river, to the northeast of Kolo Atas village (83 TO 172.3), showing the highly tectonised nature of the rocks. The clasts and matrix are both slickensided, showing both vertical and horizontal movements. The matrix consists of sheared red clay. Note the slight boudinage of the chert and calcilutite clasts.

are slickensided showing both vertical and horizontal movements (Plate 3.1B).

In the Kolo Atas area, the matrix of the melange consists of argillaceous material. The matrix is sometimes calcareous mudstone containing both planktonic and benthic foraminifera of Middle Miocene to Pliocene age. In the Poh Neck, the melange has a matrix of red scaly clay and locally, of sheared serpentinite, mixed with argillaceous marlstone containing planktonic foraminifera of Late Miocene to Pliocene age (Plate 3.2B).

The thickness of the melange is not definitely known, but the outcrops suggest that it ranges from few metres to several kilometres in different areas. In Poh Head, to the north of Balantak village, melange with a reddish brown scaly clay matrix, occurs in narrow exposure less than 50 metres wide.

In thin section, the matrix of the melanges consists of grey and yellowish siliceous argillaceous rocks, in places calcareous, containing detrital grains of quartz (65%), muscovite (30%) and minor iron oxides (e.g. 83 TO 147.2). The terrigenous quartz grains are polycrystalline, and the larger grains of silt size are monocrystalline, subhedral and angular in shape; some of them show undulatory extinction and strained features. The cryptocrystalline quartz occurs as smaller grains and in the matrix. Muscovite grains of silt size are elongate or prismatic in shape. Some of the larger grains also show wavy extinction. Most of the larger grains of muscovite and quartz show parallel alignment defining the shear surfaces or foliation of the matrix.

Where the matrix is calcareous, it is a wackestone and/or lime mudstone containing numerous planktonic foraminifera and minor benthic foraminifera of Middle Miocene to Pliocene age (e.g. 83 TO 50B). The wackestone also contains skeletal debris including coralline algae, and very minor quartz detritus (2%) with size grades up to

Plate 3. 2A Photograph of road-cutting exposure, to the north of Salodik village (Poh Neck, 83 TO 109.1), showing the foliated matrix of melange, which consists of mixed serpentinites and marlstone and contains planktonic foraminifera of Late Miocene to Pliocene age. Note the boudinaged small chips of limestone and chert and the highly fractured block of limestone in the bottom. These features are typical effects of diapiric flow under high fluid pressures.

0.2 mm. Most of the wackestone and lime mudstone matrix are pervasively sheared.

In places the melanges have a matrix of mixed serpentine and marlstone (83 TO 109.1). The serpentine is foliated and typically green in colour which is highly contrasted to the pale grey colour of the marlstone. Numerous planktonic foraminifera and minor detrital grains of quartz of silt size occur in the marlstone. The quartz grains are mostly monocrystalline and very angular in shape.

The typical scaly clay matrix is foliated and consists largely of clay and minor cryptocrystalline quartz stained by Fe-Mn oxides, which gives rise to the reddish brown colour of the matrix (e.g. 83 TO 115.2).

A. Shape of the clasts

In outcrops the melange exhibits a progressive fragmentation of the blocks through boudinaged clasts which are commonly bounded by fracture surfaces. The clasts may be modified by pervasive shearing and converted to phacoids showing delicate tails trailing away into matrix. The edges of the clasts are commonly highly fractured and fragmented to form angular chips with non-abraded features.

Subrounded to angular clasts predominantly of sedimentary rocks are found locally. They are probably detached from the disintegration of earlier olistostrome or pebbly mudstone deposits.

B. Nature of the clasts

As described previously, the melange contains various rock types including (i) ophiolitic, (ii) sedimentary, (iii) volcanic and (iv) metamorphic rocks.

(i) Ophiolite fragments

The ophiolitic fragments include peridotites, gabbro, dolerite and basalt. They are commonly very altered, and the peridotites are highly serpentinised, often showing mesh texture due to derivation of serpentine from the olivine. The serpentinised peridotites contain antigorite, showing flaky features and/or honeycomb or net structure (e.g. 83 TO 192.3). The pyroxene is invariably altered to chlorite and iron oxides. Some of the pyroxene shows marginal alteration and is replaced by hornblende and chlorite (e.g. 83 TO 7). The less altered peridotites consist largely of olivine and subsidiary hypersthene and magnetite. Chromite is present in some peridotite fragments.

Basic rock fragments show ophitic or subophitic textures. Phenocrysts of plagioclase, including labradorite (An55-68) and minor andesine (An33-48), show typical carlsbad/albite twinning, some with zoning. Most of the plagioclase is marginally or zonally altered and replaced by carbonate and sericite. Ortho- and clino-pyroxene are subhedral and prismatic in shape and are highly altered. Uralitisation of pyroxene is also seen in some rocks. Chloritisation of pyroxene is common, possibly related to hydrothermal alteration.

(ii) Sedimentary fragments

The sedimentary fragments include limestone, marlstone, quartzose arenite, quartz-rich lithic arenite, pebbly arenite, chert and calcilutite. Some fragments of quartz-rich lithic arenites, which are usually more rounded in shape than the other fragments are thought to have been derived from the pebbly mudstones of the Kolo Beds.

The limestone and marlstone fragments include grainstone, packstone and wackestone. Most of them contain both planktonic and benthic foraminifera. They range in size from pebbles up to tens of metres, and usually are angular in shape or irregular in form. Most of them are tabular or elongate and some are boudinaged.

In thin section the grainstone consists largely of skeletal grains, up to 65 % in 83 TO 50B; 83 TO 111.2, including benthic and planktonic foraminifera, echinoids, algae and molluscs. Additional grains present in small amounts, include terrigenous quartz and glauconite pellets. These grains are usually set in a matrix consisting of lime mud, micrite and iron oxides. Some of the skeletal grains are replaced by micrite and some are infilled by sparry calcite.

The packstones are compositionally identical to the grainstones, but they contain much more lime mud matrix. Skeletal grains forming less than 30% of the rock, consist largely of benthic foraminifera and subsidiary planktonic foraminifera, algae, echinoids and skeletal debris of unidentified macroinvertebrates. Although most of the skeletal debris is micritised, the internal structures of benthic foraminifera are still visible. Monocrystalline quartz detritus is present in small amounts in some marlstone fragments. Some packstones also contain glauconite pellets of silt size in very minor amounts (usually less than 2%). Matrix of the packstone usually consists of lime mud and micrite, but in some rocks minor quartz occurs in the matrix.

The wackestone may contain skeletal debris forming up to 30% of the rock, and very minor quartz detritus. Some of the wackestones are laminated. Skeletal debris consists largely of planktonic and benthic foraminifera and minor algae, echinoids and molluscs. In most cases, the benthic foraminifera have been slightly flattened, due to compression of the fragments during development of the melange.

The quartzose arenite fragments are dominated by quartz detritus (97%) and very minor muscovite (e.g. 83 TO

51). The quartz grains are well-sorted, subangular, and texturally super mature. They are wholly monocrystalline; most of them show wavy extinction and strain features. Overall, the quartz grains show an interlocking mosaic, possibly due to compaction and/or deformation during the history of development of the melange, or probably they have been lithified before inclusion in the melange.

Quartz-rich lithic arenite fragments are composed of up to 60% quartz detritus and up to 30% lithic fragments (e.g. 83 TO 41). The quartz grains are identical to these of the quartzose arenite described previously. Rock fragments include siliceous mudstone, argillite, slate, phyllite, schists and volcanics, and grade up to 1 mm across. The volcanics are highly altered, and some of the mudstones are recrystallised. Some of the argillite and mudstone contain ghosts of microfossils, probably radiolaria.

The chert fragments are typically angular in shape, with size grades up to several tens of metres long. They are composed essentially of cryptocrystalline quartz, and contain recrystallised ghosts of radiolaria. The calcilutite fragments, compositionally comprise lime mudstone and wackestone, and contain numerous calcareous microfossils set in a matrix of a mixed lime mud and clay and iron oxides. The microfossils occur as walled-calcispheres, most of which were micritised and infilled by sparry calcite.

(iii) Volcanic fragments

Rhyodacitic vitric tuff fragments occur just to the east of Kolo Atas village. They are highly fractured and jointed, and yellowish in colour due to weathering; with size grades up to several metres long.

In thin section the rocks consist of vitric tuff laminae alternating with crystal tuff laminae. The crystal tuff consists largely of crystals of plagioclase and pyroxene and subsidiary glass-supported matrix. The plagioclase consists of labradorite (An 55-68) and andesine (An34-45), subhedral and prismatic in shape with size grades up to 0.2 mm long. Most of the plagioclase grains are altered. Ortho and clino-pyroxene are present, most of them are partly altered and replaced by chlorite and iron oxides. Quartz, hornblende and biotite crystals are also present in small amounts. The vitric tuff consists predominantly of glass shards with minor microlites of plagioclase and pyroxene (e.g. 83 TO 38.4).

(iv) Metamorphic fragments

The metamorphic fragments consist largely of low-grade metasedimentary rocks, including phyllite, slate, quartzite and metachert. The metasediments show an interlocking texture of elongate crystals, giving rise to a dimensional preferred orientation (granoblastic-elongate texture). Mica-schist fragments are foliated, defined by the parallelism of long flat crystals of mica and the elongate quartz.

The clasts in the Kolokolo Melange can be readily recognised as having been detached from the adjacent tectonic elements, i.e. the overriding ophiolite suite of the subduction complex and the underplating continental margin. The clasts of chip to block size, including serpentinised peridotites, gabbro, dolerite, basalt, radiolarian red chert and metamorphic rocks are derived from the ophiolite suite. The underplating continental margin is represented by those clasts of sediments, including quartz-rich and carbonate sediments and the highly altered intermediate to acidic volcanic rocks.

3.2.4 Age determination of melange

The age of the melange is determined on the basis of two main geological factors :

1. The youngest clasts or blocks of the melange are derived from the Palaeogene Salodik Limestones. The limestone and marlstone fragments (e.g. 83 TO 50B; 83 TO 111.2) contain both benthic and planktonic foraminifera. The benthic foraminifera includes Lepidocyclina sp., and Amphistegine sp., and the planktonic foraminifera includes Orbulina universa D'Orbigny, Globorotalia sp., Globorotalia menardii D'Orbigny, Globigerina sp., Globigerinoides sp. and Sphaeroidinellopsis subdehiscens (Blow) of Oligocene to Middle Miocene age. These fragments are compositionally, texturally and biostratigraphically identical to the Salodik Limestones, and represent the youngest rocks occurring as clasts or blocks in the melanges.

2. The occurrence of planktonic foraminifera and minor benthic foraminifera, including Orbulina sp., Orbulina universa D'Orbigny, Globorotalia sp., Globorotalia menardii D'Orbigny, Globorotalia cultrata, Globorotalia tumida, Globorotalia scitula Brady, Globigerinoides immaturus Bronnimann, Globigerinoides sacculifer Brady, Globigerinoides obliquus Bolli, Globigerinoides trilobus Reuss, Globigerina venezuelana Hedberg, Globoquadrina altispira Cushman & Jarvis, Sphaeroidinella subdehiscens Blow and Sphaeroidibellopsis seminulina Blow of Middle Miocene to Pliocene age.

These two age indicators strongly suggest that the formation and development of the melange occurred since the late Middle Miocene, subsequent to the collision of the Banggai-Sula Platform with the Eastern Sulawesi Ophiolite Belt or subduction complex. The formation of melange continued into Pliocene time.

The other significance of the occurrence of planktonic foraminifera within the matrix of the melange is that the collision was developed in submarine and open-sea tectonic setting.

3.2.5 Discussion and Interpretation

In the area surrounding Kolokolo Bay, the melanges are overlain unconformably by the Neogene Kolo Beds. In the outcrops along the Kapali river, the melanges are easily distinguished from pebbly mudstone (or perhaps some of them may be called olistostrome) which form the basal succession of the Kolo Beds. The main differences of the two units is essentially due to the mechanism and genesis of the rocks. The melange typically shows the matrix intruding the clasts and commonly with pervasively sheared pelitic matrix and boudinaged clasts, whereas the pebbly mudstone shows characteristics of a sedimentary origin, such as bedding and stratification, some rocks may show fabric orientation of clasts, and abraded clasts. The olistostrome commonly has a matrix of unsheared marlstone, and clasts which are more rounded than those of the melange, and the clasts show no boudinaged features.

Although the Kolokolo Melange occurred in a submarine tectonic setting as indicated by the presence of planktonic foraminifera within the matrix, the influence of currents on sedimentation is not evident.

Origin of the Melanges

The clasts of Kolokolo Melanges can be readily recognised as having been detached from the adjacent tectonic elements. Clasts of chip to block size, including serpentinised peridotites, gabbro, dolerite, basalt, metamorphic and radiolarian chert were derived from the ophiolite suite. Those clasts of sediments including quartz-rich arenite and carbonate rocks and the highly altered intermediate to acidic volcanic rocks represent a continental margin sequence.

The quartzose arenite and quartz-rich lithic arenite were derived from the Jurassic Kapali Beds, and the reddish limestone and marlstone are detached from the Late Jurassic Sinsidik Beds and Nambo Beds. At least some of the radiolarian chert and calcilutite have come from the Late Cretaceous Luok Beds. The foraminiferal limestone and marlstone, undoubtedly, were derived from the Palaeogene Salodik Beds. The volcanic fragments are believed to have been derived from the Permo-Triassic Mangole Volcanics and Tolokibit Volcanics of Banggai-Sula Platform (Sukamto, 1975a; Surono and Sukarna, 1984; Supanjono and Haryono, 1984).

On the basis of the determinations of rock assemblages forming the melanges, structures and nature of the clasts and the matrix, and the the presence of foraminifera within the matrix of the melanges, it is apparent that there are at least three possible mechanism attributed to the origin and development of the Kolokolo Melanges, namely,

i) tectonism,
ii) diapiriic processes and
iii) olistostrome (sedimentary melange).

1) Tectonism

The mechanism is essentially overthrust, which as Hsu (1970) pointed out, like a bulldozer picks up fragments or blocks from the underlying rock units, giving rise to a mixture of fragments detached from the overriding plate with those detached from the underplated tectonic elements. At this stage, the the mixture may have been spontaneously accompanied by the introduction of a pelitic matrix, which is derived from any rock units involved in collision. The introduction and presence of pelitic matrix as discussed later, may have been initiated or even accelerated by diapiric processes.

In the case of the Kolokolo Melanges, their origin is

Plate 3.2B Photograph of Kolokolo Melange exposing up-stream in the of Kapali river, Kolo Atas area (83 TO 172.2), showing the highly sheared pelitic matrix and clasts of various rock types, including quartzose arenite, quartz-rich lithic arenite, limestones, chert and ophiolitic rocks.

Plate 3.2C Photograph of oil seep occurring in alluvial deposits at Kolo Atas village. The oil seep contains very dark or black muddy tar mixed with kerosene (very dark below the fallen trees). Similar occurrences are found in at least 5 localities in the vicinity of Kolo Atas village.

closely related to the collision of the Banggai-Sula Platform (BSP) against the Eastern Sulawesi Ophiolite Belt (ESOB), in which the ESOB overrode the BSP. In fact, the Kolokolo Melanges always occur associated with or along the Batui Thrust and Balantak Fault System, which are considered to be the surface expression of the collision of the ESOB against BSP.

The youngest rock units involved within the melanges are the Eocene to Early Miocene Salodik Limestones. The formation of the melanges therefore, must be post Middle Miocene. This is consistent with the age of foraminifera found in the matrix of the melanges, which have been dated as Late Miocene to Pliocene.

In general, foliation or shearing in the matrix and the boudinaged feature of clasts trend NE-SW in the Kolo Atas area, and in E-W direction in Poh Neck and Balantak coast. This preferred orientation is parallel or subparallel to the arcuate form of the Batui Thrust - Balantak Fault System. In places, the trends may have been disrupted by fault movements, especially post-collision strike-slip and vertical movements due to diapiric processes discussed below.

ii) Diapiric processes

During and subsequent to the collision of the Banggai-Sula Platform against the Eastern Sulawesi Ophiolite Belt, the introduction of pelitic matrix into the mixture fragments or blocks of different types of rock was essentially instrumented by a diapiric process, in which water mixed pelitic material (ductile clay and/or mudstone or marlstone matrix) was injected into the irregular spaces between the fragments, that are arranged chaotically or in a slightly subparallel boudin fashion. The pelitic material may have come from any of the sedimentary units of Banggai-Sula Platform, which contain invariably substantial amounts of pelitic material. As described previously, the collision took place in a submarine tectonic setting, and hence, the water mixed pelitic matrix was readily available to be injected by the diapiric process into the spaces between the fragments.

In many places, outcrops show that shear surfaces tends to be parallel to the edges of the fragments or blocks. Most of the edges of blocks are highly fractured and even fragmented to form the extremely angular chips showing no indication of a superficial transportation process. In many places, it is observed, that the pelitic matrix was introduced into cracks and fractures of the disintegrating fragments or blocks in the melanges. Intrusion or injection of pelitic materials is essentially due to the diapiric process, which was continuously active during the history of development of the melanges.

In the Kolo Atas area, oil seeps occur mostly in places adjacent to the outcrops of the melanges. In other places, such as the area to the north of Bungku, Kayu Merangkak area (some 50 km to the north of Kolo Atas village), and on the southern flank of the Batui Mountain Ranges, oil seeps occur within the fault zone (Simandjuntak et al., 1982, 1983; Surono et al., 1984). The occurrence of oil seeps associated with the melanges are closely related to the diapiric process. Generation of oil from the organic material which is significantly present in some rocks of Mesozoic sediments described previously in Chapter 2, may increase the mobility of shale.

On the basis of data observed in the field in Timor, Barber et al. (in press) proposed a model to explain the relationship between wrench faulting, shale diapirism and mud volcanoes (Fig. 3.3). Shale diapirism and the formation of mud volcanoes have been attributed to the following factors:

1. Sedimentary loading due to rapid sedimentation.
2. Tectonic loading due to overthrusting.
3. Abnormally high pore pressures developed in uncompacted shale formations.
4. Generation of oil and gas from organic material included in shale (cf. Hedberg, 1974).
5. Dehydration of expandable clays, through mineralogical transformation : e.g. smectite alters to illite.
6. Density inversion with light, buoyant, plastic clays beneath a denser and more rigid overburden.
7. Faulting and seismic activity in regions active tectonics.

Barber et al. (op. cit) pointed out, that all these factors may be contributory to a specific occurrence of diapirism.

Most of these factors can be readily seen to have operated in the formation of the Kolokolo Melange. Tectonic loading due to overthrusting after the collision of ophiolite belt against continental margin appears to be more important than sedimentary loading. This overthrust has caused the overpressured of shale and other argillaceous or pelitic rocks in the continental margin sequence. As described previously in Chapter 2, shale, mudstone, argillaceous rocks occur as intercalated beds in the Triassic Lemo Beds, Jurassic Kapali Beds, Late Jurassic Sinsidik Beds and Nambo Beds, Late Cretaceous Luok Beds and in Buya Formation in Banggai-Sula Islands (Surono & Sukarna, 1985; Supanjono & Haryono, 1985).

In the East Arm of Sulawesi, most of the rock units are in fault contact with the adjacent units. Some of the faults are still active or re-activated. Seismic activity (McCaffrey et al., 1982) suggests that the collision may still be in progress. Tectonic activity is believed to have initiated and accelerated the diapiric process within the Kolokolo Melanges.

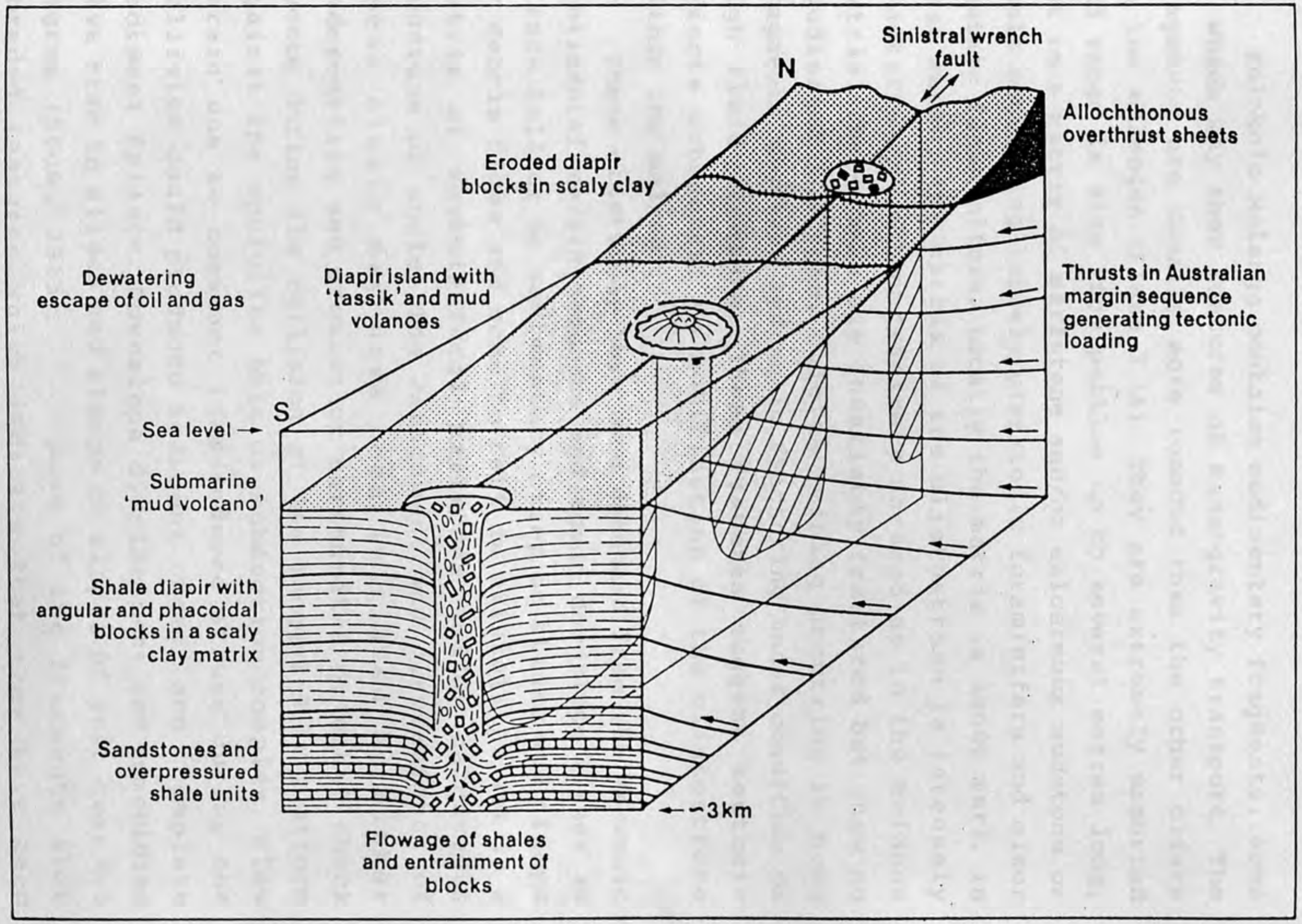


Fig. 3. 3 Conceptual model of the relationship between shear zones, shale diapirs and mud volcanoes based on the seismic reflection and field observations (After Barber et al., in press)

iii) **Olistostrome**

Kolokolo Melange contains sedimentary fragments, some of which may show features of mass-gravity transport. The fragments are usually more rounded than the other clasts in the melanges (Plate 3.3A). They are extremely unsorted and range in size from pebbles up to several metres long, set in a matrix of marlstone and/or calcareous mudstone or shale containing largely planktonic foraminifera and minor benthic foraminifera. Locally the matrix is sandy marl. In most cases, the matrix of the olistostrome is intensely fractured but not pervasively sheared as in the melange matrix. The clasts are invariably fractured but show no boudinaged features. Calcite veining occurring in most fragments suggests hydraulic fracturing under condition of high fluid pressure. These features suggest tectonic effects subsequent to mobilisation of the olistostrome within the melange.

These clasts may have been detached from the Mesozoic continental margin sequence and moved downslope either as a rock fall or by sedimentary creep or slides and slumps or debris flows and come to rest as olistoliths set in a matrix of foraminiferal marlstone and/or calcareous mudstone or shale. The detachment and displacement of these clasts may have been initiated by either undercutting and erosion or earthquakes or other shock events during the collision of the Banggai-Sula Platform against the ophiolite belt or subduction complex. Slow strain due to constant load-induced stress during the collision could produced sediment creep, and complete sediment failure. Downslope displacement and remoulding give rise to slides and slumps on slopes of just over 0.5 degree (Stow, 1985). Some of the fragments show abraded features which indicate that they have been transported and deposited by mass-movement of gravity flows, by which the sediments moved downslopes and are

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Plate 3.3A Photograph of Kolokolo Melange in a road-cutting exposure, to the south of Poh village (83 TO 111.3), showing a highly fractured matrix of marlstone rich in planktonic foraminifera of Middle Miocene to Pliocene age. Note the subrounded clasts of quartz-rich arenite and limestone. Some of the clasts show abraded features suggesting that they may have originated as olistostrome or pebbly mudstone.

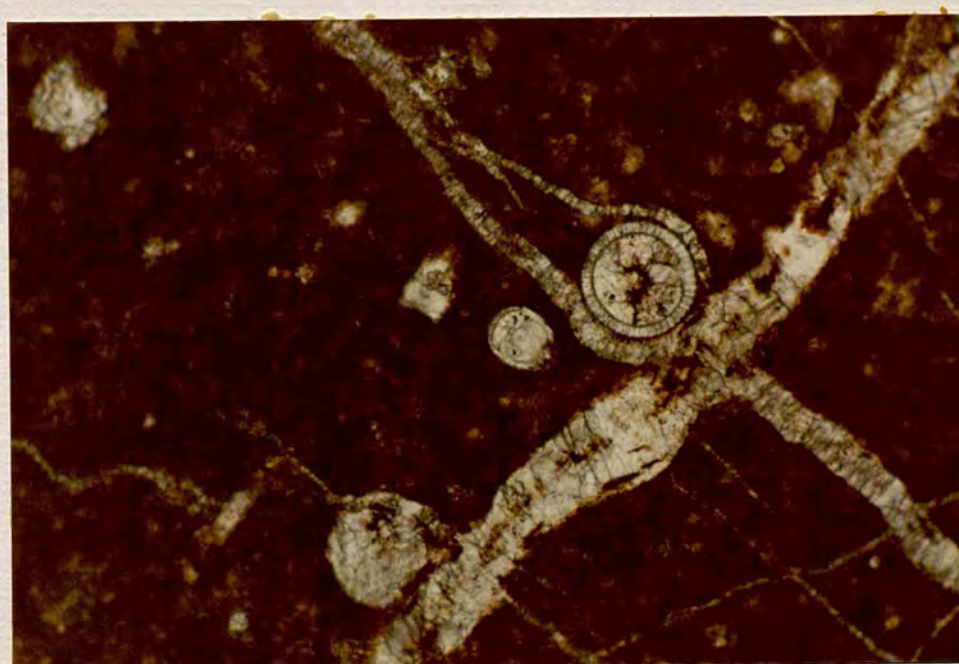


Plate 3.3B Photomicrograph of calcilutite occurring as fragments in the Kolokolo Melange (83T054), showing walled radiolaria and highly fractured lime mudstone. Plane polarised light, 40X.

driven by gravitional forces from shallower to deeper water (Middleton and Hampton, 1976; Saxov and Nieuwenhuis, 1982; Hein, 1982; Hill et al., 1984; Gorsline, 1984).

Kolokolo Melanges have been formed since and during collision of the Banggai Sula Platform against the ophiolite belt. This coincided with the formation of olistostromes, some of which were mobilised and included as clasts or fragments within the Kolokolo Melanges.

3.3 THE BALANTAK OPHIOLITE

3.3.1 DEFINITION

The ultramafic and mafic rocks with the associated radiolarian chert and minor metamorphic rocks occurring in the northern half of Poh Head, East Arm of Sulawesi are informally named the Balantak Ophiolite. The term ophiolite is used in this study following the definition established by the Penrose Field Conference, Russia (1972).

3.3.2 DESCRIPTION

The Balantak Ophiolite which outcrops over more than half of the East Arm of Sulawesi, occurs largely in the western part of the East Arm and the northern part of Poh Head. The distribution of the ophiolite is shown on the geological map prepared by the Geological Research and Development Centre (GRDC), Geological Survey of Indonesia (Fig. 3.1). It is everywhere in fault contact with Mesozoic to Palaeogene sediments and in many places is covered by Neogene post orogenic coarse clastic sediments and volcanogenic turbidites (Batui Group) and Quaternary coralline reefs (Simandjuntak et al.,1983; Surono et al.,1984; Rusmana et al.,1984). In the East Arm, the ophiolite occurs in an arcuate belt trending in an east-west direction, and convex towards the north-northwest, 270 km long with a maximum width of exposure of some 70 km in the western part of the East Arm. On closer field observation, the ophiolite is highly tectonised; faults and thrusts repeatedly occur within the ophiolite.

The ophiolite sequence is best developed in Poh Head, where it consists of a narrow zone of serpentinised peridotite, gabbro, anorthosite, trondhjemite, norite, sheeted dolerite, pillow basalt and hyaloclastite (Fig.

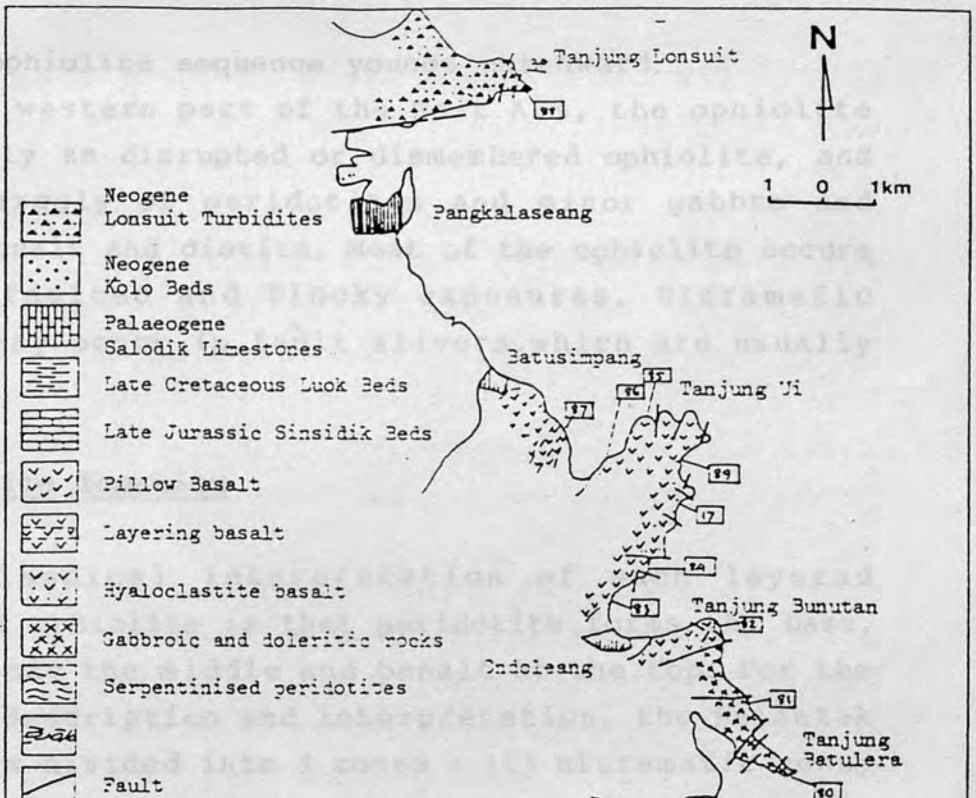


Fig. 3.4 Geological traverse map of the Balantak-Tanjung Lonsuit coast, showing the occurrence of Balantak Ophiolite, which is dominated by mafic rocks and very minor serpentinised peridotites.

3.4). This ophiolite sequence youngs northward.

In the western part of the East Arm, the ophiolite occurs mostly as disrupted or dismembered ophiolite, and consists largely of peridotites and minor gabbro and dolerite, basalt and diorite. Most of the ophiolite occurs in highly faulted and blocky exposures. Ultramafic sequences may occur in fault slivers,which are usually imbricated.

A. Ophiolite Zonation

The classical interpretation of such layered sequences of ophiolite is that peridotite forms the base, gabbroic rocks the middle and basalt at the top. For the purpose of description and interpretation, the Balantak Ophiolite is divided into 4 zones : (l) ultramafic zone, (ii) gabbro zone, (iii) sheeted dyke zone and (iv) basalt zone. In most cases, the contact between units are not clearly seen due primarily to the highly faulted nature of the rocks. Repetition due to imbrication of the rocks results in uncertainty of thickness of the units.

(i) Ultramafic Zone

Ultramafic rocks occur abundantly over most of the area occupied by the Balantak Ophiolite, particularly in the western part of the East Arm of Sulawesi. In Poh Head, however, serpentinized peridotites occur only in a narrow zone less than 50 m thick. The main body of the ultramafic rocks occurs further to the west in the Bunta area. The northern part of the Batui Mountain Range consists largely of ultramafic and minor mafic rocks, which are bounded by the Batui Thrust-Balantak Fault System with the Salodik Limestones to the south.

The ultramafic zone consists predominantly of noncumulate peridotites, consisting largely of harzburgite and subsidiary dunite. Suria-Atmadja et al. (1972) described the occurrence of wehrlite, lherzolite and pyroxenite in Central Sulawesi further to the west. Cumulate peridotites occur locally, usually in fault-bounded exposures, and their relationship with the noncumulate peridotites is not clearly defined. In the Ampana river, south of Ampana township however,small exposures show cumulate peridotites in association with overlying cumulate gabbroic rocks (Rusmana et al., 1984). Most of the ultramafic rocks are serpentinised to varying degree.

On the Balantak coast, just to the north of Tanah Merah, the ultramafic zone consists wholly of serpentinized harzburgite, which is dark green with a waxy appearance, slickensided in hand specimen, and brown on weathered exposures. Most outcrops show shearing. In hand specimen the ultramafic rocks are pale-green and vitreous. Locally, they are sheared or foliated, sometimes are folded.

In thin section, the harzburgite generally shows a xenoblastic granular texture and consists of olivine, pyroxene and minor chromium spinel and iron ores. Olivine crystals may be up to 15 mm long, but average 2-4 mm. They are usually subhedral with slightly corroded boundaries. Some of the olivine grains show kink banding, and as a result original grain boundaries become obscure. In most rocks, the olivine grains are partly altered to serpentine and show typical net or honeycomb or mesh-structures (Plate 3.4A, B). In some grains, alteration is accompanied by rounded patches of magnetite, which tend to occupy the corroded interiors. Olivine and/or its alteration product may be form up to 70% of the rocks (e.g. 83 TO 193.4).

Orthopyroxene is present in the harzburgite, and ranges in size from 2-4 mm, and is subhedral and prismatic in shape. Some of the pyroxene occurs as exsolution lamellae. Some of the pyroxene also shows kink banding.

Plate 3.4A Photomicrograph of serpentinised harzburgite from the Balantak Ophiolite occurring on the coast of Kolokolo Bay (i.e. 83 TO 195.2), showing mesh structure due to alteration of olivine to serpentine. the presence of flaky antigorite along the margin of the olivine. Crossed polars, 40X.

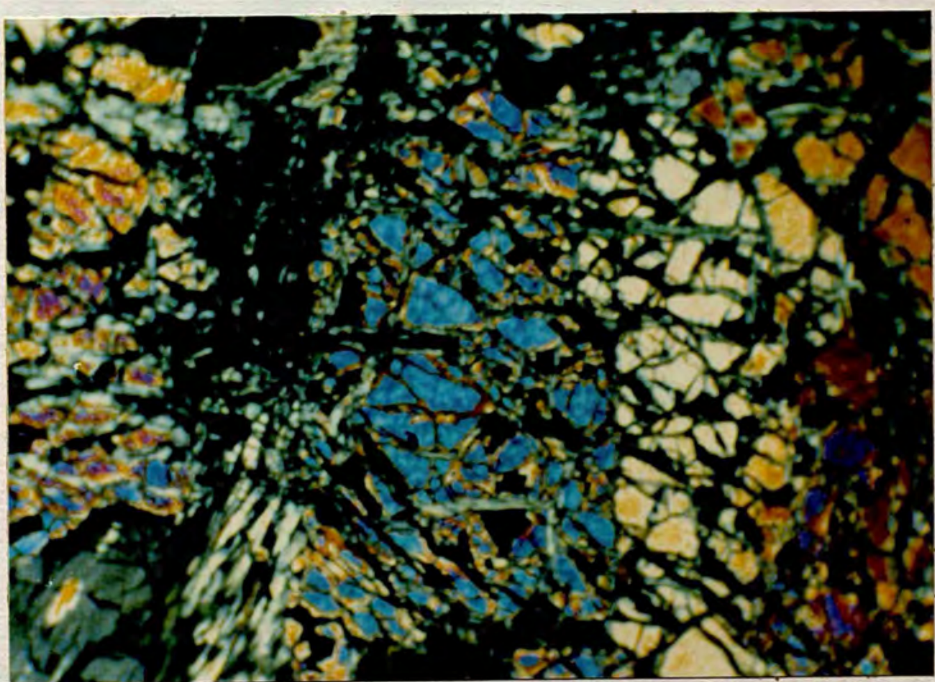


Plate 3.4B Photomicrograph of serpentinised dunite from the Balantak Ophiolite occurring on the coast of Kolokolo Bay (i.e. 83 TO 195.11), showing the honeycomb structure due to alteration of olivine to serpentine. Crossed polars, 40X.

Pyroxene may constitute up to 20% of the rocks (e.g.83 TO 193.3). Chromite occurs as intergranular euhedra and as inclusions within olivine, and generally forms grains up to 1 mm long. Magnetite occurs in small amounts as intergranular euhedra and as inclusions within the olivine and pyroxene. Fibrous chrysotile and platy antigorite occur in most harzburgite, as alteration products of olivine. The grains are interlocking anhedra with varying degrees of deformation.

The dunite exhibits a xenoblastic granular texture and consists predominantly of olivine, and minor accessory chromite. Grain size is generally more variable than that in harzburgite,and ranges from 2-10 mm, but occasionally large crystals of olivine up to 5 cm long may be present. Olivine may constitute up to 95 % of the rocks (e.g.83 TO 193.1). The pyroxene grains are subhedral with size ranging from 0.3-3 mm, and present in very small amounts. Chromium spinel occurs as intergranular euhedra and as inclusions within the olivine, and generally forms grains up to 1 mm long. Small magnetite inclusions within olivine appear to have developed as a by product. Kink banding is present in some of the dunites, resulting in very diffuse boundaries.

(ii) Gabbroic Zone

Gabbroic rocks occur in many places in the East Arm of Sulawesi. The best exposures of the gabbroic zone occur in Poh Head along the northern side of the Balantak Fault System (Fig. 3.1; Fig. 3.4). In some places, the gabbroic zone is underlain by a thin zone of serpentinized peridotites. In the western part of the East Arm, the gabbroic zone occurs in an irregular pattern, due to the highly tectonised nature of the bodies, giving rise to a disrupted or dismembered ophiolite suite.

On the coast north of Balantak, the gabbroic zone is

Plate 3.5A Photograph of exposure on the coast of Tanjung Batang, to the north of Balantak village (i.e. 83 TO 71), showing highly fractured and faulted gabbroic rocks of the Balantak Ophiolite. Note the pegmatitic gabbro in the left.

Plate3.5B Photograph of exposure on the coast of Tanjung Bunutan to the north of Balantak village (i.e. 83TO89), showing highly sheared gabbroic rocks in the Balantak Ophiolite.

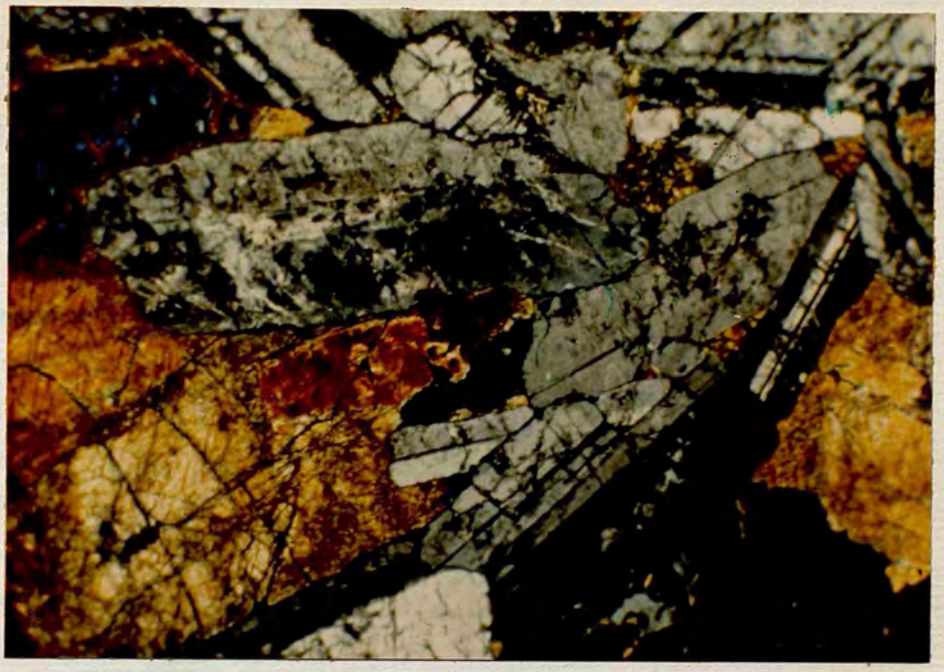


Plate 3.5C Photomicrograph of gabbro from the Balantak Ophiolite occurring in Dua Island, to the north of Balantak village (i.e. 83 TO 72.1), showing hypidiomorphic granular texture. The plagioclase and pyroxene crystals are both highly fractured, and are partially altered. Crossed polars, 40X.

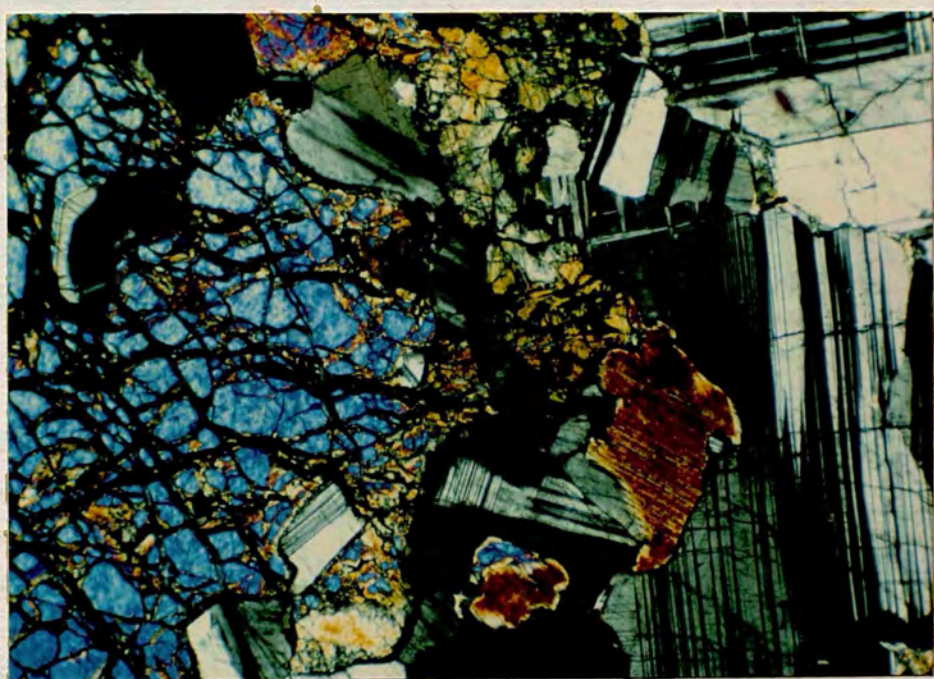


Plate 3.5D Photomicrograph of olivine gabbro from the Balantak Ophiolite occurring along the road to Siuna, to the east of Salodik village (i.e. 83 TO 12.1), showing polysynthetic twinned plagioclase and typically fractured olivine. Hypersthene slightly interlocks with plagioclase. Crossed polars, 40X.

in fault contact with the sheeted dyke zone,which is also in fault contact with the pillow basalt zone. In the Balantak section, the gabbroic zone is made up several lithological types, including granular gabbro, cumulate gabbro, homogeneous gabbro and gabbro pegmatite.

The granular gabbro includes a wide variety of gabbroic rocks, which in thin section are characterised by allotriomorphic and/or hypidiomorphic granular textures. Some of them, especially those occurring the shear zones, may show a degree of planar or linear fabric. The gabbros are composed of plagioclase, hypersthene, augite, olivine, chlorite and very minor opaque iron oxides. The plagioclase is labradorite (An 54-64), subhedral and prismatic or tabular, with size grades up to 4 mm, showing polysynthetic twinning, and may constitute up to 50% of the rock (e.g. 83 TO 74.2). Most of plagioclase crystals are altered, marginally or zonally, and replaced by sericite, carbonate, clay, zeolite and albite.

The pyroxene consists largely of augite and subsidiary hypersthene. It is grey greenish, subhedral and prismatic in shape with crystals up to 3 cm long. The hypersthene is typically highly pleochroic in green to red and has parallel extinction which contrasts with the oblique extinction of the augite. Up to 30% pyroxene may be present in the rocks (e.g.83 TO 74.2). Some of the pyroxene crystals,especially along their margin, are altered and replaced by hornblende and uralite (e.g. 83 TO 7). In some gabbros, the pyroxene is highly chloritised,suggesting hydrothermal alteration along faults which are frequently seen within the rocks. Some rocks contain up to 15% of olivine showing typical mesh-structure due to alteration to serpentine (e.g. 83 TO 12.1).

The other varieties of gabbro occurring in the zone include norite and troctolite. The norites are holocrystalline with hypidiomorphic granular texture and

Plate 3.6A Photograph of coastal exposure near Ondoleang village, to the north of Balantak village (i.e. 83 TO 83), showing a sheared zone within the doleritic rocks of the Balantak Ophiolite.

Plate 3.6B Photograph of exposure at coast of Teku, to the north of Balantak village (i.e. 83 TO 80.3), showing pegmatitic gabbro in the Balantak Ophiolite. Shearing and segregation also developed in this rock.

are composed of plagioclase (labradorite) up to 60% and hypersthene up to 35% (e.g.83 TO 115B). Augite is present in small amounts (5%) in some norites (e.g. 83 TO 75).

The plagioclase crystals are subhedral and tabular in shape, range in size from 1-4 mm, and show typical albite twin lamellae, in some cases, combined with carlsbad and/or pericline twinning. Some of the plagioclase is schillerized; some may be altered to zoisite, epidote and sericite.

Hypersthene is green, subhedral and prismatic in shape, and grades up to 3 mm long. Some of the hypersthene contains inclusions of augite, showing stronger birefringence than the host mineral. Small inclusions of magnetite are also present in some crystals of hypersthene. A fibrous amphibole occurs particularly along or around the periphery of hypersthene crystals, due to uralitisation of the pyroxene (e.g.83 TO 115B).

Troctolite consists essentially of plagioclase and olivine,and very minor pyroxene with a hypidiomorphic granular texture (Plate 3.7A). The plagioclase consists of labradorite-bytownite (An 60-70), subhedral and tabular in shape, ranging in size from 0.3-2 mm, and may constitute up to 75 % of the rocks. Olivine crystals are subhedral to anhedral with size grades up to 1.8 mm across, and show typical net or honeycomb structures with microfractures radiating out into the olivine. Some of the olivine crystals show corona structure, being rimmed with reaction products consisting largely of orthopyroxene and minor amphibole. Up to 25% olivine may be present in the rocks (e.g.83 TO 12.2).

In places, the gabbroic zone also contains anorthosite and trondhjemite. They are typically light in colour and occur as thin layers or sheets several centimetres to tens of centimetres thick. Anorthosite consists of nearly pure plagioclase with only very minor amounts of pyroxene and iron oxides. The plagioclase

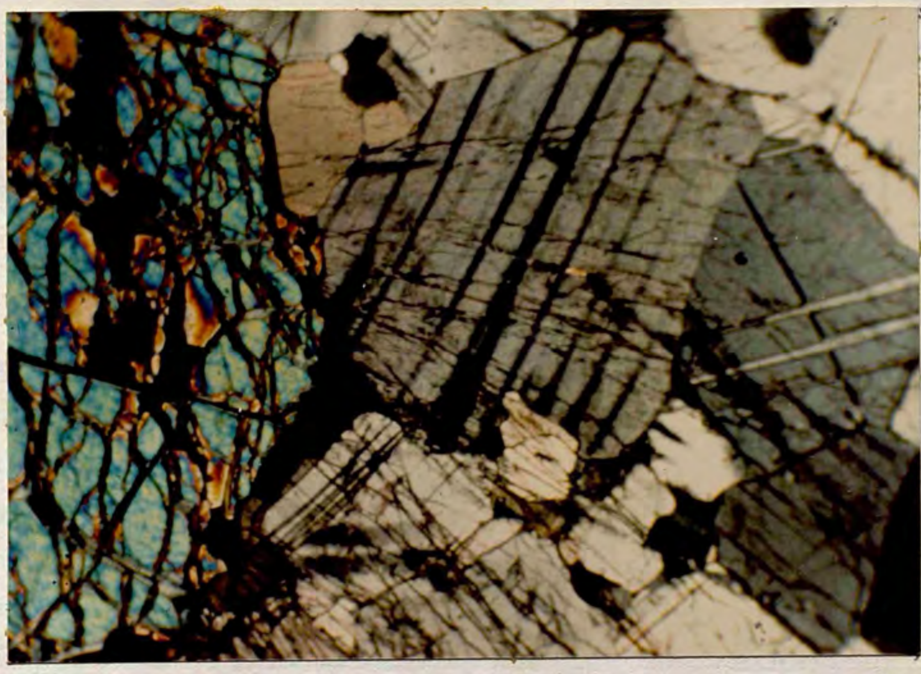


PLate 3.7A Photomicrograph of troctolite from the Balantak Ophiolite occurring in a bouldery exposure to the east of Salodik village (i.e. 83 TO 12.2), showing highly fractured olivine and polysynthetic twinned plagioclase. Crossed polars, 40X.

Plate 3.7B Photograph of exposure on the coast of Tanjung Padingkian, to the north of Balantak (i.e. 83 TO 87), showing a doleritic dyke (dark). Note the development of criss-cross fractures filled by calcite and zeolite throughout the rocks.

consists of labradorite-bytownite (An 66-70); subhedral and tabular in shape with size grades up to 3 mm long. Some of the plagioclase crystals are marginally altered and replaced by sericite and carbonate.

(iii) Sheeted Dyke Zone

In the East Arm of Sulawesi, the sheeted dyke zone occurs in discontinuous exposures along the coast of Kampangar-Tanjung Batang, north of Balantak (Fig. 3.2). The outcrops suggest that, structurally the sheeted dyke zone occurs above the gabbroic zone and underlies the pillow basalt zone. The boundaries are not clearly defined due to the highly tectonised nature of the rocks.

In general the dykes are randomly oriented. However, on closer field observation, a significant number of dykes trend in a NW-SE direction, with variable steep dips to NE and/or SW. The irregular pattern of dyke attitudes might have been caused by faults and thrusts, which are seen frequently disrupting the rocks. Silver et al.,(1983) recognised NNW-SSE trending sheeted dykes along the tributaries of the Binsil river.

In outcrops the dykes show a variation in colour from dark to darker green and they are commonly chilled against one another. The dykes range in thickness from 5 cm to 1 m, and up to tens of metres in length. They are fine grain with doleritic and/or porphyritic texture. Field relationships in Tanjung Batang reveals that in the lower part of the sheeted dykes, most of the dykes are in fault contact with the gabbroic zone. The internal part of the sheeted dykes may show multiple and subparallel individual dykes. Brecciation is frequently observed within the dyke swarms. Williams and Malpas (1972) suggested that this brecciation results from a late stage igneous gas action or fluidization.

The upper contact between the sheeted dykes and the

Plate 3.8A Photograph of an outcrop of the Balantak Ophiolite on the coast of Teluk Ui (i.e. 83 TO 86), showing layered basalt dipping moderately toward northwest.

Plate 3.8B Photograph of an exposure of the Balantak Ophiolite in a river cliff to the east of Pon (i.e. 83 TO 10), showing the occurrence of very thin cherty calcilutite filling the interstices between the pillows.

pillow basalts is also not clearly seen due to the highly tectonised nature of the rocks. Williams and Malpas (1972) described a sheeted dyke swarm extending well-up into the overlying volcanics in Bay of Newfoundland, Canada. In Poh Head however, as will be discussed later, the basalt zone consists mostly of pillow lavas with some layered basalt and hyaloclastite but no dykes have been seen.

The sheeted dyke zone, as in the other parts of the ophiolite belt, is highly faulted (Fig. 3.4) and so its thickness is difficult to define. On the basis of a profile crossing the zone, the thickness of the sheeted dyke may be up to 3 km. The estimation is based on profile perpendicular to the layering of basalt along the coast of Tanjung Bola to Kampangar. Thickening of this rocks is must be due to repetition or duplication by thrusting.

In thin sections, the dykes consist of fine to medium grained, grey to greenish dolerite showing typical intergranular texture. They are composed essentially of plagioclase, clinopyroxene, and opaque iron oxides. In some rocks, brown hornblende may be present, probably developed at the expense of pyroxene.

The plagioclase consists largely of labradorite (An 55-60) and subsidiary andesine (An22-30), which is subhedral and tabular and/or prismatic in shape, with size grades from 0.2-1 mm long. Some of the plagioclase shows zoning and polysynthetic twinning. The plagioclase may constitute up to 65 % of the rocks (e.g. 83 TO 87). Most of the plagioclase crystals are altered, marginally or zonally. Some of them are totally altered and replaced by carbonate, sericite, zeolite, and albite. Calcic plagioclase has partly broken down to cloudy or dirty albite (e.g. 83 TO 84.2).

Clinopyroxene occurs in the interstices between the tabular plagioclase crystals, as pale brown anhedral grains, and may form up to 40% of the rocks (e.g. 83TO87). Most of the clinopyroxene grains are highly chloritized, some are uralitized and replaced by a fibrous green actinolite (e.g. 83T017B). Hornblende occurs in small amounts in some rocks. Some of the hornblende may be a late magmatic phase, but it mostly represents low-grade thermal alteration of primary clinopyroxene. Opaque iron oxides include magnetite and ilmeno-magnetite, occurring as common accessory minerals. Some of them may be derived from alteration of clinopyroxene.

Coleman (1972) considered that the occurrence of sheeted dykes is due to the rising of magma along a single fracture. The magma is chilled against a previously emplaced and solidified dyke. Each successive dyke is emplaced within the middle of previously emplaced dykes. In Poh Head, this succession is not clearly observed. The exposures show that dykes were intruded after the crystallisation and solidification of an earlier gabbroic magma.

(iv) Basalt Zone

In the East Arm of Sulawesi,the basalt zone consists largely of pillow lavas and subsidiary massive and layered lavas, hyaloclastite and minor brecciated basalt. Calcilutite is locally present associated with the basalt zone, usually filling the interstices between the pillows.

In an exposure in a steep ciff to the west of Poh village, the pillow lavas face-up towards northwest (Plate 3.8A, B; 3.9A, B). Basaltic lavas are well-exposed along the coast of Tanjung Bunutan-Batusimpang to the south of Tanjung Lonsuit (Fig.3.4) and along the Bombon river in Poh Head. In the Bombon river, the pillow basalt is topping northwestwards (Plate 3.9B). In the coast to the east of Batusimpang, Balantak area, the pillows face-up and young northwards.

In the western part of the East Arm, the basaltic rocks usually occur in an imbricated complex and

Plate 3.9A Photograph of an exposure of the Balantak Ophiolite in a river cliff to the east of Pon (i.e. 83 TO 10), showing pillow basalt. The pillow lavas are facing-upwards toward the north-northwest direction.

PLate 3.9B Photograph of an exposure of the Balantak Ophiolite in the Bombon River, to the south of Binsil village (i.e. 83 TO 136.4), showing pillow basalt younging towards the north-northwest.

associated with ultramafic and gabbroic rocks. Small and isolated exposures of basaltic rocks occur throughout the East Arm of Sulawesi (Fig.3.1).

In Tanjung Bunutan (e.g. 83 TO 89) the outcrops suggest that the basal portion of the basalt zone merges into the sheeted dyke swarm in a gradual fashion, with the number of dykes increasing downwards. The upper contact is not seen in this area, but in the Bombon river (e.g. 83 TO 137.1) lensoidal calcilutite occurs on top of pillow lavas and/or hyaloclastite, marking the transition from igneous extrusion (i.e. the basalt zone) to a sedimentary regime (see description in the other part of this chapter). The calcilutite lenses range in thickness from 5 to 20 cm and up to 2 m long (Plate 3.13A). The sedimentary regime includes the volcanigenic Lonsuit Turbidites, which unconformably overlie the pillow lavas. In tributaries of the Binsil river, the calcilutite and Lonsuit Turbidites both contain Late Miocene to Pliocene planktonic foraminifera, indicating an open marine environment. In the western part of the East Arm, however, the pillow lavas are associated with radiolarian chert and calcilutite. In Poh Head, there is much less pelagic sediments associated with the basalt zone than in the western part of the East Arm of Sulawesi.

The basalts are dark-grey, greenish and brownish on weathered surfaces. The lower portion of the basalt zone consists of massive and layered basaltic rocks,which are invariably dipping towards the north-northwest. Most of the pillow basalt and hyaloclastite basalt show amygdales, which are filled by secondary zeolite and/or calcite. Rounded or oval zeolite-filled amygdales up to 2.5 cm across occur in many places. The amygdales suggest that the basalts were extruded in a shallow marine tectonic setting. Zeolite and calcite also occur in the criss-crossing fractures which are commonly developed throughout the basalt zone.

Like the other ophiolitic rocks, the basalt zone is highly faulted and so its thickness is difficult to determine. The basalt occurring in Poh Head uniformly dips at 30 degrees towards north-northwest, assuming no repetition on thrust planes, the thickness of the basalt zone may account for at least 3 km.

In thin sections, the basalts are porphyritic and intergranular with phenocrysts of plagioclase and pyroxene up to 2 mm long and a groundmass containing microlites of plagioclase and pyroxene (Plate 3.10). The plagioclase consists of labradorite (An 56-70), subhedral and tabular or prismatic in shape, ranging in size from 0.2-2 mm long, often shows polysynthetic twinning, and some with carlsbad-albite twinning. Most of the plagioclase crystals are partially altered, marginally or zonally, and replaced by sericite, zeolite, carbonate, clay and some by albite (e.g. 83 TO 71.1; 83 TO 76). The plagioclase may constitute up to 60% of the rocks (e.g. 83 TO 10).

Pyroxene consists largely of clinopyroxene (i.e. augite) which is usually subhedral and prismatic in shape, ranging in size from 0.2-0.8 mm long. Most of the pyroxene grains are partly altered to chlorite, epidote, iron oxides and locally by hornblende. In some rocks, however, the pyroxene is mostly chloritised (e.g. 83 TO 71.1), due to hydrothermal alteration along the fault zone, from which the rocks were sampled. Up to 40% of pyroxene may be present in the rocks (e.g. 83TO76). Magnetite occurs as intergranular euhedra and as inclusions within the olivine, and has probably developed as a by product of serpentinization.

In the Bombon River, the pillow basalts show chilled margins which range in thickness from 0.5-5 cm. The chilled rims consist largely of glassy material and may contain variolites or microlites indicating incipient crystallisation.

The basalts are dark-grey and greenish in thin

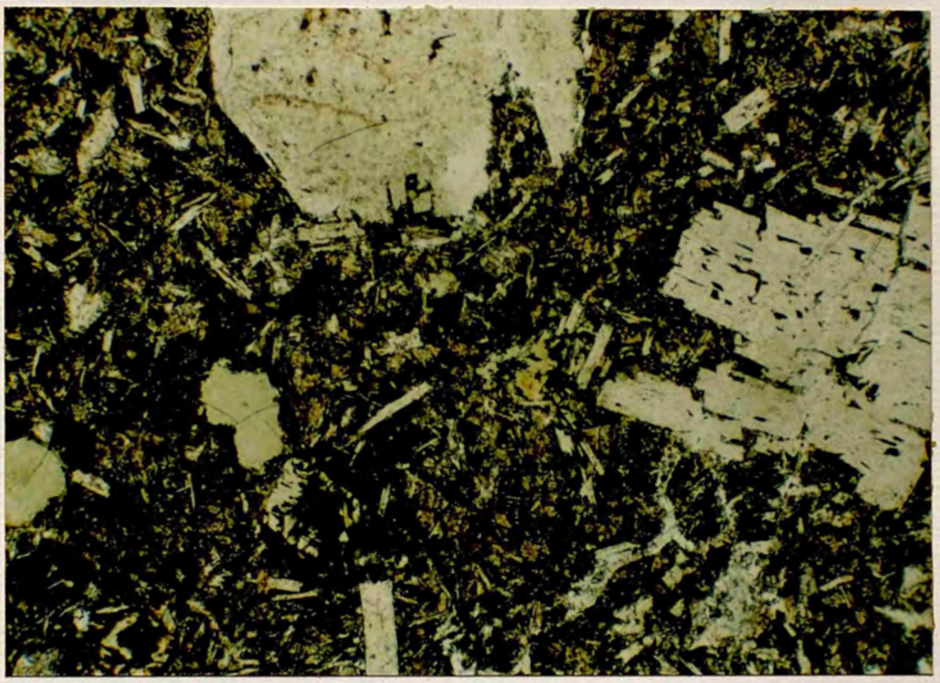


Plate 3.10 Photomicrograph of basalt from the Balantak Ophiolite occurring as pillow lavas in the cliff to the east of Poh (i.e. 83 TO 10), showing porphyritic texture with a phenocryst of plagioclase and a matrix consisting of microlites,plagioclase and pyroxene. The pyroxene is partially altered and replaced by chlorite and iron oxides. The plagioclase is also partially altered and replaced by sericite, zeolite and carbonate. Plane polarised light, 40X.

section; most of them are highly altered. They are porphyritic with phenocrysts of plagioclase and pyroxene and a groundmass of glassy material containing microlites of plagioclase, pyroxene and magnetite (e.g. 83 TO 134.3). In some rocks, volcanic glass which may constitute up to 20% (e.g. 83 TO 134.3) is partially devitrified and replaced by clay, albite and iron oxides.

3.3.3 BOBA BEDS

A. Definition:

Radiolarian chert-dominated sediments occurring in the Boba river are informally named the Boba Beds (Fig. 3.6). The basal contact is not seen at the type locality, but a small exposure on the coast to the west of Kolo Bay suggests that the unit was deposited on top of the ophiolite suite.

B. Synonymy

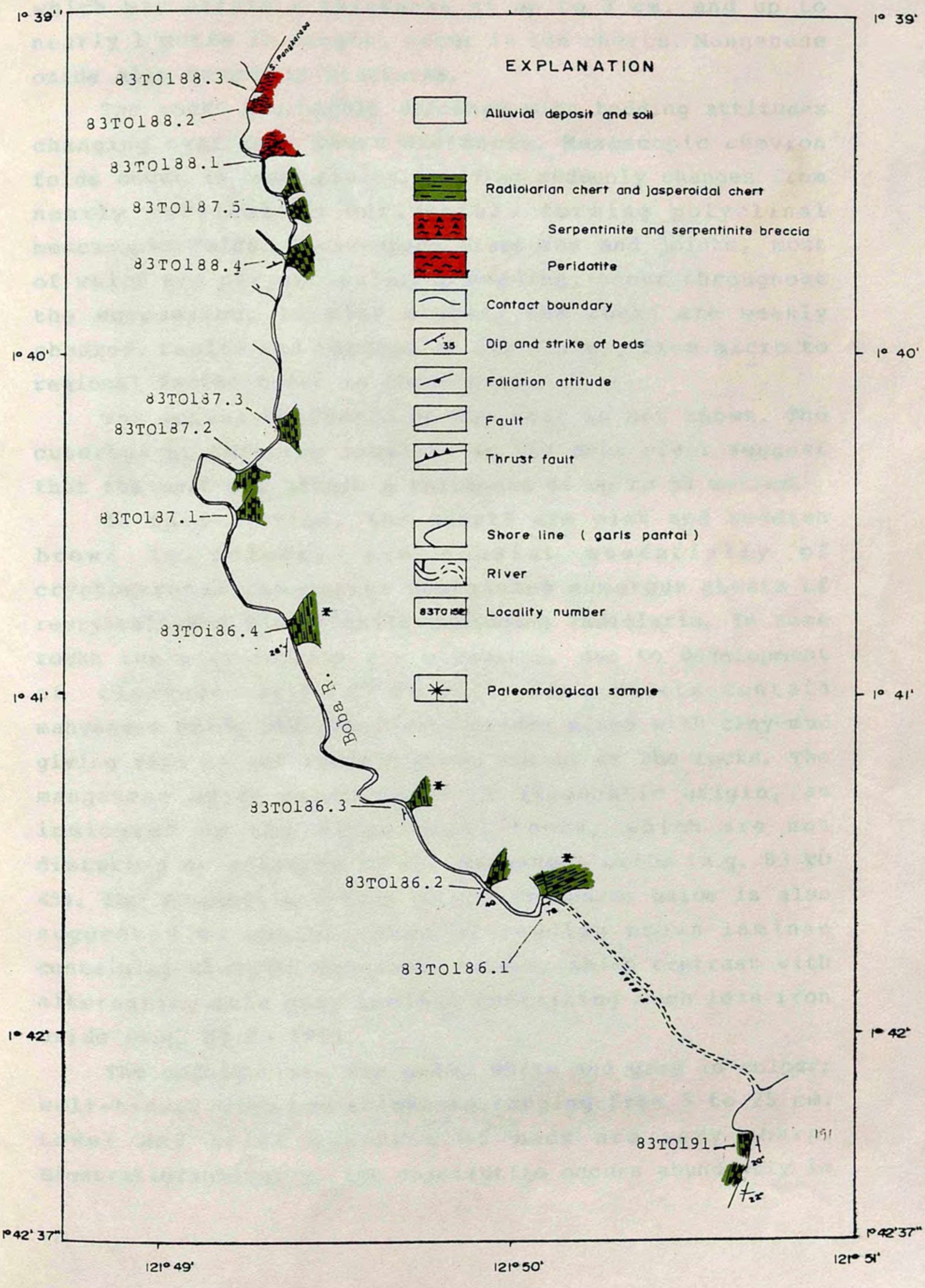
The Boba Beds are included within the Matano Formation (Simandjuntak et al., 1983).

C. Description

The Boba beds occur in blocky or discontinuous exposures in the Boba river, along the coast near Boba village, and west of Kolo Bay. The unit also occurs in fault-bounded exposures on a small hill to the north of Kolo Atas village, and as small exposures in many places associated with the ophiolite in the western part of the East Arm of Sulawesi. The chert has not been found in Poh Head.

The Boba Beds consist largely of radiolarian chert. Calcilutite which commonly occur together with the cherts in the Boba area (western part of the East Arm) are also included in the Boba Beds. The stratigraphic relationship between the chert and calcilutite is not clearly known. The cherts are pink, light-grey, greenish and red in colour, thinly bedded with bed thickness ranging from 2 - 10 cm. Basal and upper contact of each bed is sharply defined. Thin parallel laminae (usually less than 2 cm thick) occur in some beds of chert. Manganese oxide in the

Fig. 3.5 Geological traverse map of Boba River, showing the occurrence of Boba Beds.



form of irregular lenses, slablike or flattened nodules, which may attain a thickness of up to 3 cm, and up to nearly 1 metre in length, occur in the cherts. Manganese oxide also occurs in fractures.

The rocks are highly deformed with bedding attitudes changing over very short distances. Mesoscopic chevron folds occur in many places. Bedding suddenly changes from nearly vertical to horizontal, forming polyclinal mesoscopic folds. Close-space fractures and joints, most of which are perpendicular to bedding, occur throughout the succession. In many places, the rocks are weakly sheared. Faults and thrusts on all scales, from micro to regional faults occur in these rocks.

The actual thickness of the unit is not known. The outcrops at the type locality in the Boba river suggest that the unit may attain a thickness of up to 50 metres.

In thin section, the cherts are pink and reddish brown in colour, and consist essentially of cryptocrystalline quartz containing numerous ghosts of recrystallised microfossils including radiolaria. In some rocks the microfossils are elongated, due to development of cleavage (e.g. 83 TO 161). Some cherts contain manganese oxide and other iron oxides mixed with clay-mud giving rise to the reddish brown colour of the rocks. The manganese oxide seems to be of syngenetic origin, as indicated by the microfossil tests, which are not disturbed or coloured by the manganese oxide (e.g. 83 TO 45). The syngenetic origin of the manganese oxide is also suggested by the presence of reddish brown laminae containing abundant manganese oxides, which contrast with alternating pale grey laminae containing much less iron oxide (e.g. 83 TO 191).

The calcilutites are pale, white and grey in colour; well-bedded with bed thickness ranging from 5 to 25 cm. Lower and upper contacts of beds are very sharp. Biostratigraphically, the calcilutite occurs abundantly in

Plate 3.11A Photograph of an exposure of the Boba Beds on a hill to the north of Kolo Atas village (i.e. 83TO173.1), showing a sequence dominated by well-bedded red chert. Note the highly fractured rocks.

Plate 3.11B Photograph of an outcrop of the Boba Beds in the Boba river (i.e. 83TO186.1), showing sheared zone and highly fractured rocks. Pink calcilutite is also present.

the upper part of the succession. Their contact with the succession of chert is not seen in the Boba River section.

In thin section, the calcilutites are pale-grey and yellowish lime mudstone, and consist primarily of micrite containing numerous calcispheres of microfossils. Most of the microfossils are walled-calcispheres, which are filled by micrite and some by sparry calcite (e.g. 83 TO 26). There are no calcispheres filled or replaced by siliceous or cryptocrystalline quartz.

Thin parallel laminae from several millimetres up to 2 cm thick, occurring in some calcilutite beds, are defined by an alternation of calcisphere-supported wackestone and lime mudstone laminae. The calcisphere-supported wackestone laminae are cemented by micrite mixed with iron oxides resulting in a darker-brown colour, while the lime mudstone laminae are calcite-cemented and light-grey in colour. In some rocks, pressure solution cleavage is weakly developed. The cleavage may cut through the calcispheres.

D. Calcilutite in Poh Head

In Poh Head, the calcilutite occurs, locally, filling interstices between the pillows. The calcilutite is usually in the form of single layer, ranging in thickness from 0.5 to 10 cm, and up to 1 metre in length (Plate 3.13A). The calcilutite is never observed occurring in a succession of bedded pelagic sediments.

The calcilutite usually occurs associated with the amygdaloidal pillow basalts of the Balantak Ophiolite. Chert is not found associated with the calcilutite. Due to the very minor and small occurrences, the calcilutite can not be assigned as a rock unit. For the purpose of description and interpretation, the calcilutite is included in this section.

In thin section the calcilutite consists largely of

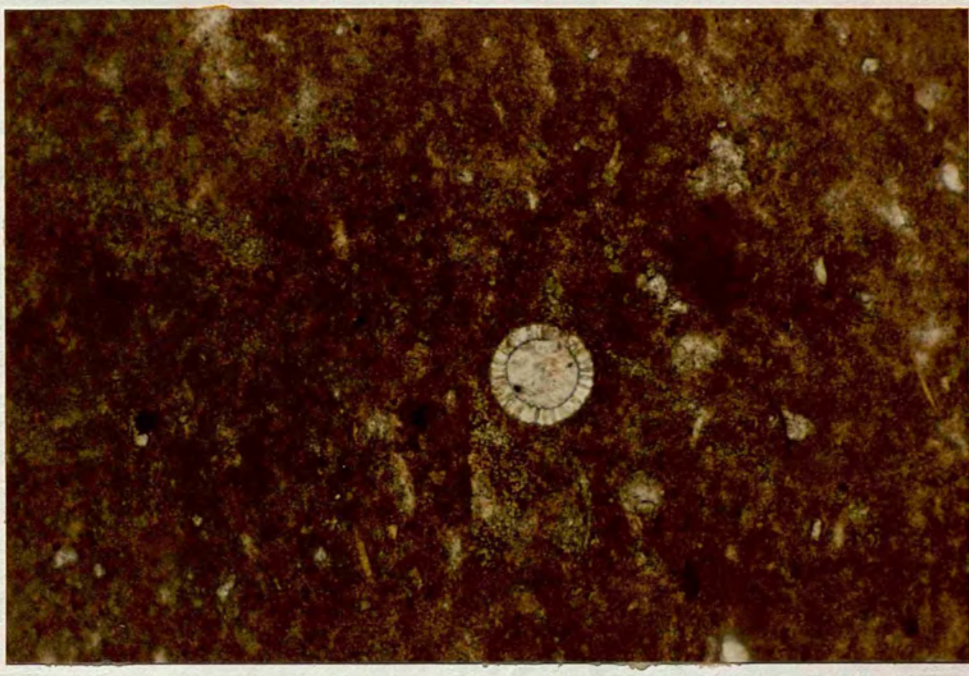


Plate 3.12 Photomicrograph of calcilutite from the Boba Beds in the Boba river (i.e. 83TO186.1), showing an walled radiolaria. The rock consists entirely of lime mud. Plane polarised light, 40X.

micritised calcite and minor iron oxides. There are no fossils or shell fragments or skeletal debris (e.g. 83 TO 78.3). Texturally this rock is typically a pelagic deposit (Plate 3.13B). The absence of chert and microfossils and unbedded nature of this rock make the calcilutite quite different from the calcilutite in the Boba Beds.

E. Biostratigraphy

The chert and calcilutite of the Boba Beds, both contain microfossils. The chert collected from Kapali River (e.g. 83 TO 50; 83 TO 173.1) contains abundant but variably preserved radiolaria, including Thanarla conica, Zipondium, Archaeodictyomitra sp., A. apiaria, Pseudodictyomitra sp. cf., P. cosmoconica, Acanthocircus sp. aff. A. multidentatus, cryptocephalic and crypthoracic nasellarians of Early Cretaceous (Valanginian) to Late Cretaceous (Early Cenomanian) identified by Dr. Benita Murchey, USGS California (person. comm., 1985), while the calcilutites contain calcareous microfossils including Globotruncana sp., Rotaliapora sp. and Heterohelix sp. of Late cretaceous age (e.g. 83 TO 26, 83 TO 35.3, 83 TO 35.4 identified by Purnamaningsih and Dr.Darwin Kadar, GRDC).

No fossils have been found in the calcilutite lenses occurring in association with the pillow lavas on Poh Head. On the basis of occurrence and stratigraphical position, the maximum age of the calcilutite in Poh Head is the same as that of the associated pillow basalt, i.e. Early Eocene to Early Oligi/cene.

F. Stratigraphic relationship

In the western part of the East Arm of Sulawesi, cherts and overlying calcilutites occur associated with the ophiolitic rocks in the Boba river. Similar rocks are also found associated with the ophiolite in many places in

Plate 3.13A Photograph of an exposure of the Balantak Ophiolite on the Bombon River (i.e. 83 TO 143), showing calcilutite filling the interstices between the pillows.

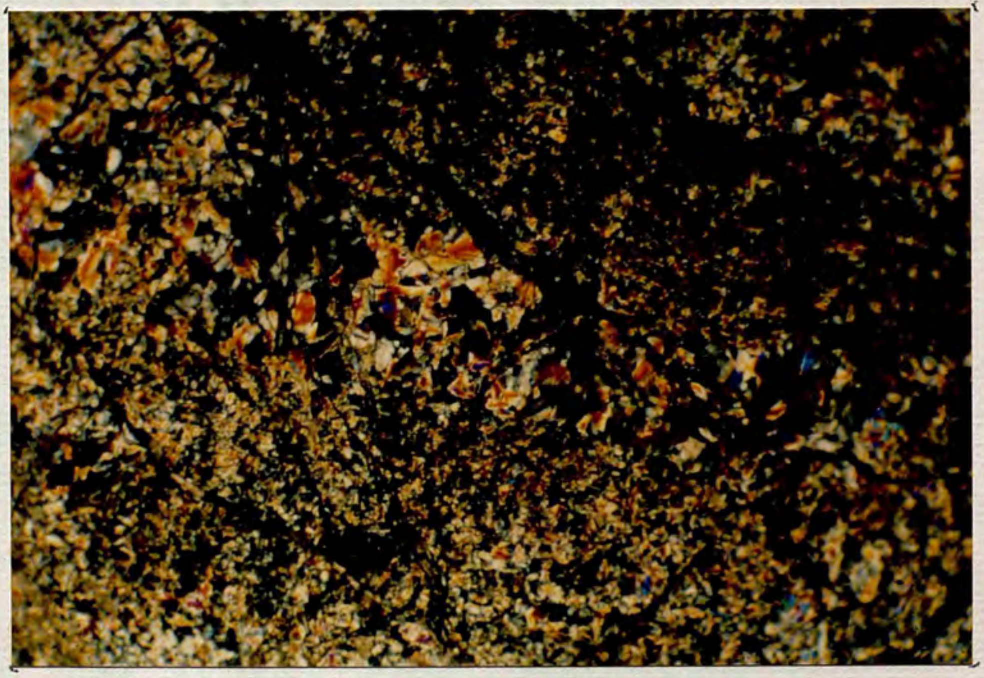


Plate 3.13B Photomicrograph of calcilutite filling the interstices between the pillow basalt on the coast just to the north of Kampangar village (i.e. 83 TO 78.3), showing that the rock consists largely of lime mud with no fossils. Calcite veinlet cuts the calcilutite. Crossed polars, 40X.

Central Sulawesi (Simandjuntak et al., 1981, 1982), in the Southeast Arm, Kabaena Island (Simandjuntak and Surono, 1983) and in Buton Island (Suharsono, et al., 1976; Smith, 1982). In Poh Head, however, only very minor and small exposures of calcilutite occur locally associated with the pillow lavas. The calcilutites usually occur in the form of lenses less than 10 cm thick and less than 2 metres long and filling the spaces (or interstices) between the pillows.

In the western part of the East Arm (i.e. Boba area), due to the highly tectonised nature of the rocks, the stratigraphical relationship of the chert and calcilutite is not definitely known. The biostratigraphy of these rocks, however, suggests that the cherts were deposited in the Early Cretaceous, which is prior to accumulation of the calcilutite in Late Cretaceous.

In Poh Head, only very minor and small exposures of calcilutite occur locally within the pillow lavas. The occurrence and stratigraphical position of the calcilutite suggest that this rock was deposited subsequent to formation of the pillow basalt since Early Eocene time. Hence, the calcilutite in Poh Head is relatively younger than that in the western part of the East Arm.

G. Discussion and Interpretation

These very fine grained sediments associated with the ophiolitic rocks, are texturally and compositionally, typical pelagic deposits consisting largely of chert and subsidiary calcilutite. The sediments are even bedded, with thin parallel laminae occurring in some beds.

Lithological association, sedimentological features and biostratigraphy of the pelagic deposits show that these rocks occurring in the western part of the East Arm (i.e. Boba area) are slightly different from those occurring in Poh Head. In the Boba area, the sequence consists largely of even-bedded chert of Early to Late Cretaceous and subsidiary bedded calcilutite of Late Cretaceous, while in Poh Head the pelagic deposit is dominated by non-fossiliferous calcilutite occurring as thin single layers or filling the interstices between the pillows of Early Eocene to Early Oligocene age.

The succession dominated by chert with abundant siliceous microfossils in the Boba Beds, was deposited below the CCD. Van Andel (1975) showed that the CCD was in the range of 3200 to 4000 metres for the Pacific and Indian Oceans during Cretaceous to Palaeocene times. The cherts in the Boba Beds, therefore, were accumulated in the Early Cretaceous at the minimum possible depth of 4000 metres.

The syngenetic manganese oxides are considered to have resulted from enrichment of oceanic bottom waters in manganese from basalts and rift-zone fluids (Blatt, 1982), which are produced by leaching from convection cells of recirculating sea water and juvenile fluids at or close to an active spreading ridge (Leeder, 1982).

The occurrence of syngenetic manganese oxides in the laminae in some beds of chert, suggests that the sediments were deposited and redistributed by the action of infrequent bottom currents, which are rich in manganese oxides. Manganese oxides occurring in irregular slab-like or nodule-like masses, probably developed diagenetically, but those filling fractures have a post-depositional origin.

The calcilutite in the Boba area is always associated with the older succession of chert, while in Poh Head, it always occurs associated with the pillow basalts. These stratigraphical features suggest that the calcilutite in the western part of the East Arm was deposited transitionally on top of the chert and the calcilutite in Poh Head was deposited directly on top of the pillow basalts.

Biostratigraphically, the calcilutite occurring associated with the pillow basalt of the Balantak Ophiolite in Poh Head is younger than calcilutite associated with chert in Boba area. This feature suggests that the chert-dominated pelagic rocks in the western part of the East Arm were deposited at a depth below CCD in ocean (basin) floored by oceanic crust. Alternatively, the calcilutite in Poh Head was deposited on top of seamounts, which form subsequent to formation of oceanic crust. Depositional setting of the calcilutite in an oceanic crust studded with seamounts becomes shallower (i.e. above CCD) toward the end of Cretaceous through the Early Tertiary times. This feature fits the younger age of the calcilutite in Poh Head than the pelagics in the Boba area. This is also suggested clearly by the absence of chert associated with the pillow basalt in Poh Head.

The third possibility, is the calcilutite in Poh Head was deposited in basin (ocean) which is floored by oceanic crust with seamounts in the region closer to the spreading axis than the oceanic crust which is associated with the chert-dominated pelagic rocks in the Boba area. This assumption fits the younger age of the calcilutite in Poh Head than the pelagics in the Boba area, and the northward younging of the Balantak Ophiolite, which is consistent with the northward younging of the oceanic crust in the Banda Sea and the inferred spreading axis to the north of Sulawesi (Chao & McCabe, in press; Lapouille, 1985). The eastward displacement along the Balantak Fault System of the Balantak Ophiolite suggests the original site of the oceanic crust was in the region to the north-northwest (See Chapter 4).

Recent geophysical study of the Banda Sea (Lapouille et al., 1985; Chao-Shing & McCabe, in press) show that the Banda Sea is floored by oceanic crust of Early Cretaceous age. The ophiolitic rocks in the East Arm of Sulawesi are interpreted to be part of the Banda Sea crust emplaced during the collision of the Banggai-Sula Platform (BSP) and the Eastern Sulawesi Ophiolite Belt (ESOB).

Scholl et al. (1968) estimate a very slow rate of accumulation of pelagic deposits (i.e. 0.01 mm/yr.) in the Chile Trench. To provide chert of tens or possibly up to more than 100 m thick such as that of Boba Beds, it requires some tens of million years for its accumulation. Deposition of the siliceous pelagic sediments of the Boba Beds on top of the oceanic crust, commenced in the Early Cretaceous.

Comparison with the chert nodules in Luok Beds

The well-bedded chert and the presence of syngenetic manganese in the Boba Beds make this unit quite distinctive from the chert nodules in the Luok Beds. Compositionally and genetically, the two units are also quite different. The cherts in Boba Beds are composed, essentially, of siliceous microfossils, while the chert nodules in Luok Beds were formed from crystallisation of chemically unstable amorphous silica within calcilutite host rocks. The depositional and tectonic setting is also quite distinctive; the cherts of the Boba Beds were accumulated at bathyal depth below CCD, while the chert nodules in the Luok Beds were formed diagenetically within the calcilutite deposited at depth above CCD. The Luok Beds occur as part of carbonate dominated-continental margin sequence, while the Boba Beds were formed as part of the pelagic cover of the ophiolite suite.

3.3.4 Age of the ophiolite rocks

An attempt has been made to determine the age of the ophiolite rocks by using radiometric analyses of the mafic rocks (i.e. gabbro, dolerite, basalt) and biostratigraphic analyses of the radiolarian chert and calcilutite which occur in association with the ophiolitic rocks. Nineteen samples of mafic rocks from the ophiolite belt were collected from different localities for radiometric analyses, 12 samples of chert and 7 samples of calcilutite for paleontological analyses.

Potassium analyses was carried out by Mr. Jerry Ingram, Department of Geology, Chelsea College, University of London. K/Ar analyses and age calculations were conducted by Dr. N.Snelling, British Geological Survey (BGS) and paleontological analyses and age determination of the radiolaria by Dr. Bonita Murchey, United States Geological Survey (USGS), California and Purnamaningsih-Siregar and Dr. Darwin Kadar, Geological Research and Development Centre (GRDC), Bandung Indonesia. The age of the peridotites cannot be obtained due to their very low potassium content.

Six samples of mafic rocks collected from the Luok-Batusimpang coast, range, in age from 93.36 $\pm$ 2.27 my to 48.13 $\pm$ 2.67 my or Late Cenomanian to Early Eocene. The age of gabbro occurring just to the south of Poh village (i.e. 83TO115) is 59.32 + 4.21 my or Late Paleocene. The basalt collected from Tanjung Ui-Batusimpang coast, Bombon River and the cliff to the east of Poh village range in age from 53.52 $\pm$ 1.48 to 32.20 $\pm$ 7.88 my or Early Eocene to Early Oligocene (Fig. 3.6).

The age of the mafic rocks which represent the upper portion of the Balantak Ophiolite suite, therefore, ranges from Late Cenomanian to Eocene (93.36 $\pm$ 2.27 to 37.73 $\pm$ 0.91 my). The wide age range of these rocks suggests that either the Balantak Ophiolite represents a substantial

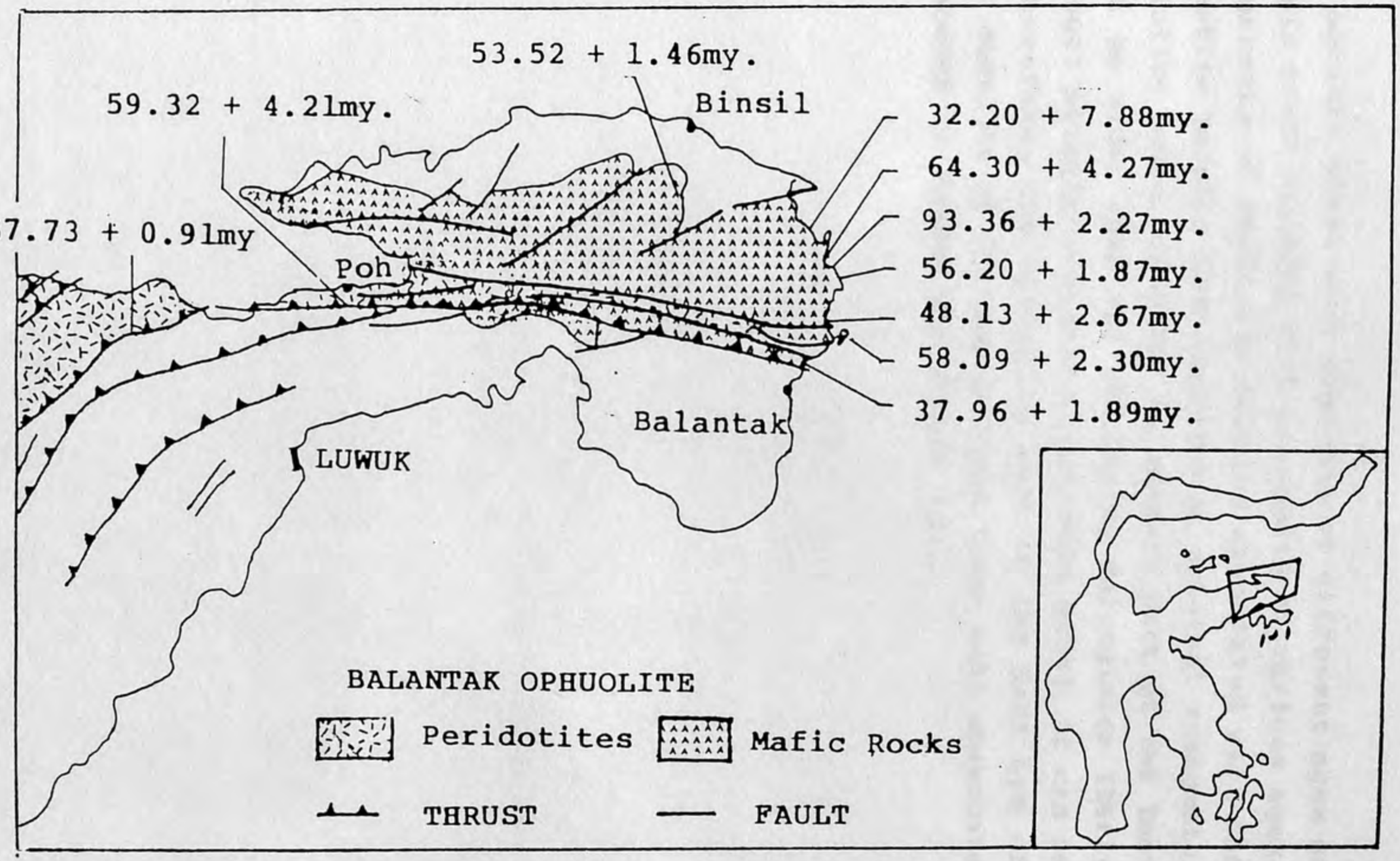


FIG. 3.6 Map showing structural configuration of Poh Head, East Arm of Sulawesi and age of the ophiolitic rocks.

area of oceanic crust with segments of different ages or an oceanic crust studded with seamounts of various ages. The occurrence of vesicular basalts associated with the hyaloclastite basalts also suggests an original seamounts. The ophiolite suite occurring in western part of the East Arm must be older than the overlaying Valanginian (Early Cretaceous) pelagic sediments (i.e. Boba Beds). It can be said, therefore, the ophiolite belt in the East Arm of Sulawesi consists of Cretaceous ocean floor with seamounts of Cretaceous to Eocene age (Table 3.1).

Fig. 3.7 Age evidence for the ophiolitic rocks.

Basalt zone — Concordant K-Ar ages on whole-rock and plagioclase from 4 samples indicate Early Eocene to Early Oligocene age (53.52 ± 1.46 to 32.20 ± 7.88 my)

Gabbro zone — Concordant K-Ar ages on whole-rock and plagioclase from 6 samples indicate Cenomanian to Early Eocene age (96.36 ± 2.27 to 48.13 ± 2.69 my).

Pelagic sediments: In the western part of East Arm, the Boba Beds contain chert with radiolaria, including *Thanarla conica*, *Zipondium*, *Archaeodistyomitra sp*, *A. apiaria* and *Pseudodictyomitra sp.* of Valanginian (Early Cretaceous) to Late Cretaceous age and the calcilutite contains *Globotruncana sp.* and *Heterohelix sp.* of Late Cretaceous age.

3.3.5 THE EASTERN SULAWESI OPHIOLITE BELT (ESOB)

The Balantak Ophiolite forms the northern portion of the Eastern Sulawesi Ophiolite Belt (ESOB). The ESOB was originally in the form of an arcuate belt, convex towards the west. The configuration of the belt has been greatly disrupted and modified by faults and thrusts, which occurred repeatedly subsequent to the emplacement of the ophiolite.

The ESOB stretches continuously from Poh Head in the north to the eastern part of Central Sulawesi and the Southeast Arm and terminates in the islands of Buton and Kabaena in the south. It is, however, discontinuous in the southern part of the Southeast Arm and Buton island (Sukamto,1975a; Smith,1983; Simandjuntak and Surono,1983), probably due to post-collision subsidence of this region.

The belt attains a maximum width of 70 km in the western part of the East Arm (Fig. 3.6). Although the belt, now, is not really continuous, but physiographically and tectonically is considered to be a belt, amounting to at least 1000 km in length. The belt is characterised by high relief with mountain ranges containing peaks nearly 3000 m high. On the basis of Bouguer anomalies across Central Sulawesi from west to east, Silver et al.,(1978) estimated that the maximum thickness amounted to at least 15 km in the western part of the belt, coinciding with the Bouguer high. They also suggest that the belt thickens westwards, reflecting the highly imbricated nature of the ophiolitic rocks in the western part. In the East Arm, the belt contains a complete ophiolite sequence, which youngs northward away from the Banggai-Sula Platform.

In the Morowali area (Central Sulawesi) the ophiolite occurs in imbricated exposures younging northwestward, away from the Banggai-Sula Platform (Simandjuntak et al.,1983). In the southern part of the belt, in Kabaena Island, the ophiolite also occurs in an imbricated

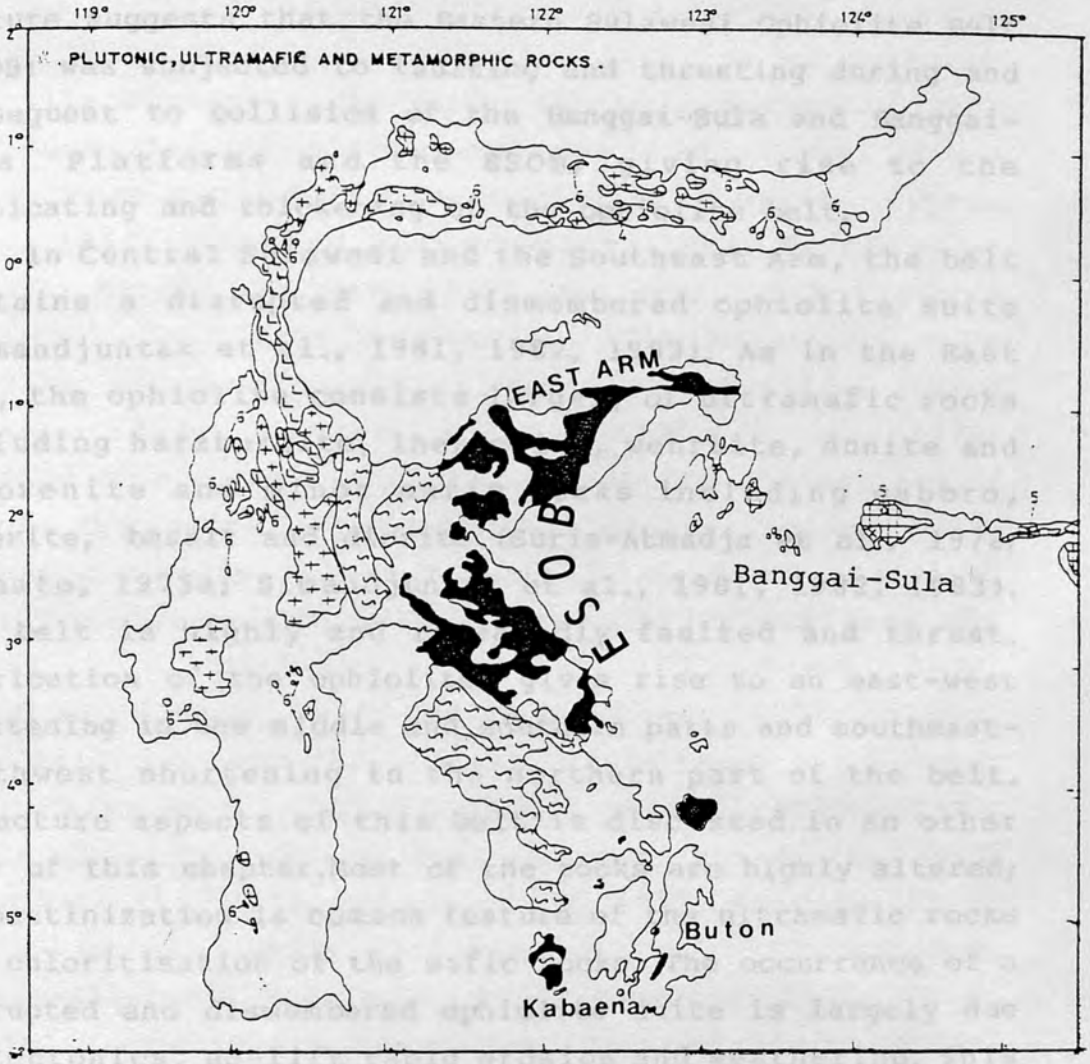
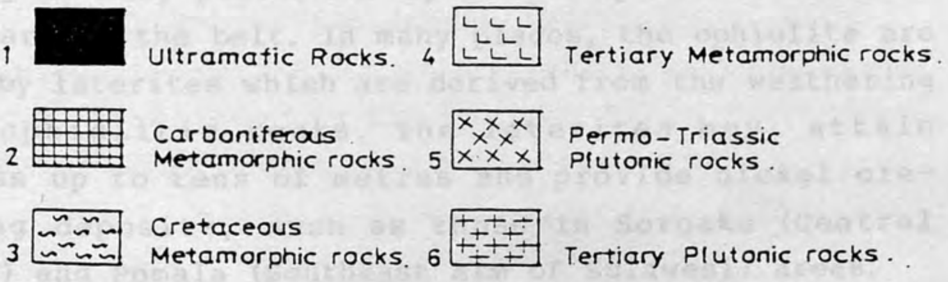


Fig. 3.8 Map showing the occurrence and distribution of the Eastern Sulawesi Ophiolite Belt (ESOB).



sequence, and youngs westwards away from the Tukang Besi-Buton Platform (Simandjuntak and Surono, 1984). This feature suggests that the Eastern Sulawesi Ophiolite Belt (ESOB) was subjected to faulting and thrusting during and subsequent to collision of the Banggai-Sula and Banggai-Sula Platforms and the ESOB, giving rise to the duplicating and thickening of the ophiolite belt.

In Central Sulawesi and the Southeast Arm, the belt contains a disrupted and dismembered ophiolite suite (Simandjuntak et al., 1981, 1982, 1983). As in the East Arm, the ophiolite consists largely of ultramafic rocks including harzburgite, lherzolite, wehrlite, dunite and pyroxenite and minor mafic rocks including gabbro, dolerite, basalt and diorite (Suria-Atmadja et al., 1972; Sukamto, 1975a; Simandjuntak et al., 1981, 1982, 1983). The belt is highly and repeatedly faulted and thrust. Imbrication of the ophiolites gives rise to an east-west shortening in the middle and southern parts and southeast-northwest shortening in the northern part of the belt. Structure aspects of this belt is discussed in an other part of this chapter.Most of the rocks are highly altered; serpentinization is common feature of the ultramafic rocks and chloritisation of the mafic rocks. The occurrence of a disrupted and dismembered ophiolite suite is largely due to tectonics, up-lift rapid erosion and weathering. This is clearly seen in the presence of mafic-dominated fragments of post orogenic coarse clastic sediments occurring in many places on top of the ophiolite rocks in middle part of the belt. In many places, the ophiolite are covered by laterites which are derived from the weathering of the ophiolitic rocks. The laterites may attain thickness up to tens of metres and provide nickel ore-producing deposits, such as those in Soroako (Central Sulawesi) and Pomala (Southeast Arm of Sulawesi) areas.

3.3.6 DISCUSSION AND INTERPRETATION

The lithological association and structural features of the Balantak Ophiolite in the East Arm of Sulawesi generally correspon to those the well-documented and well-described ophiolite complexes in other part of the world, such as the Troodos Ophiolite Complex, Cyprus (Gass, 1963; Gass and Masson-Smith, 1963; Peterman et al.,1971; Moores and Vine, 1971; Greenbaum, 1972; Lapierre and Parrot, 1972; Gass and Smewing, 1973; Miyashiro, 1973; Magarits and Taylor, 1974; Menzies and Allen, 1974; Spooner, 1974; Smewing, 1975; Smewing et al., 1975; Robinson et al., 1983; Malpas and Langdon, 1984), Bay of Islands Ophiolite, Newfoundland, Canada (Smith, 1958; Rodgers and Neals, 1963; Stevens, 1970; Dewey and Bird, 1971; Williams, 1971,1973; Church, 1972; Williams and Malpas, 1972; Williams and Smyth, 1973; Dewey, 1974; Church and Riccio, 1977; Karson and Dewey, 1978; Malpas, 1979; Salisbury and Christensen, 1978; Casey and Karson, 1981; Casey et al., 1981; Elthon et al., 1982, 1984), Semail Ophiolite Complex, Oman (Allemann and Peters, 1972; Glennie et al., 1974; Hopson et al., 1981; Smewing, 1980; Smewing et al., 1983; Alabaster et al., 1983; Gregory, 1984; Pallister, 1984; Hall, 1984; Browning, 1984), Papua New Guinea Ophiolite Complexes (Dow and Davies, 1964; Davies, 1968, 1971; Davies and Smith, 1971; England and Davies, 1973; Milsom, 1973, 1981, 1984; Rod, 1974; Coleman, 1977; Jaques, 1981; Davies, 1982; Davies and Hutchinson, 1982; Davies and Jaques, 1984).

The characteristics of an ophiolite sequence were defined and established in the Penrose Field Conference (1972), as follows, from base to top (Fig. 3.9):

E- Sedimentary section, typically including ribbon cherts, thin shale interbeds, and minor limestones.

D - Mafic volcanic complex, commonly pillows (sodic felsic intrusive and extrusive rocks may also be present)

C - Mafic sheeted dyke complex

B- Gabbroic complex, ordinarily with cumulus textures, commonly containing cumulus peridotites and usually less deformed than the ultramafic complex.

A - Ultramafic complex, consisting of variable proportions of harzburgite, lherzolite, and dunite, usually with a metamorphic fabric. The complex is more or less serpentinised. Podiform bodies of chromite may be present.

The Balantak Ophiolite which forms the northern portion of the Eastern Sulawesi Ophiolite Belt (ESOB) comprises a typical ophiolite suite as indicating by presence of a well-developed ophiolite stratigraphy consisting of two petrogenetically distinct components, i.e. an ultramafic component in the lower part and a "magmatic component" in the upper part. In Poh Head, the Balantak Ophiolite contains ultramafic rocks occurring in narrow exposures (less than 50 m wide). In Poh Neck, the ultramafic rocks occur in exposures 300 m wide, (Fig.2.2). The ultramafic component may attain a thickness of more than 5 km in the Bunta-Ampana area, to the west of the Poh Head. As will be described in Chapter 4, displacement and separation of ultramafic rocks is essentially due to dextral movement of Poh Head and Ampana-Bunta area the Balantak Fault System of at least 150 km.

The ultramafic component consists largely of harzburgite and subsidiary dunite, lherzolite, pyroxenite,

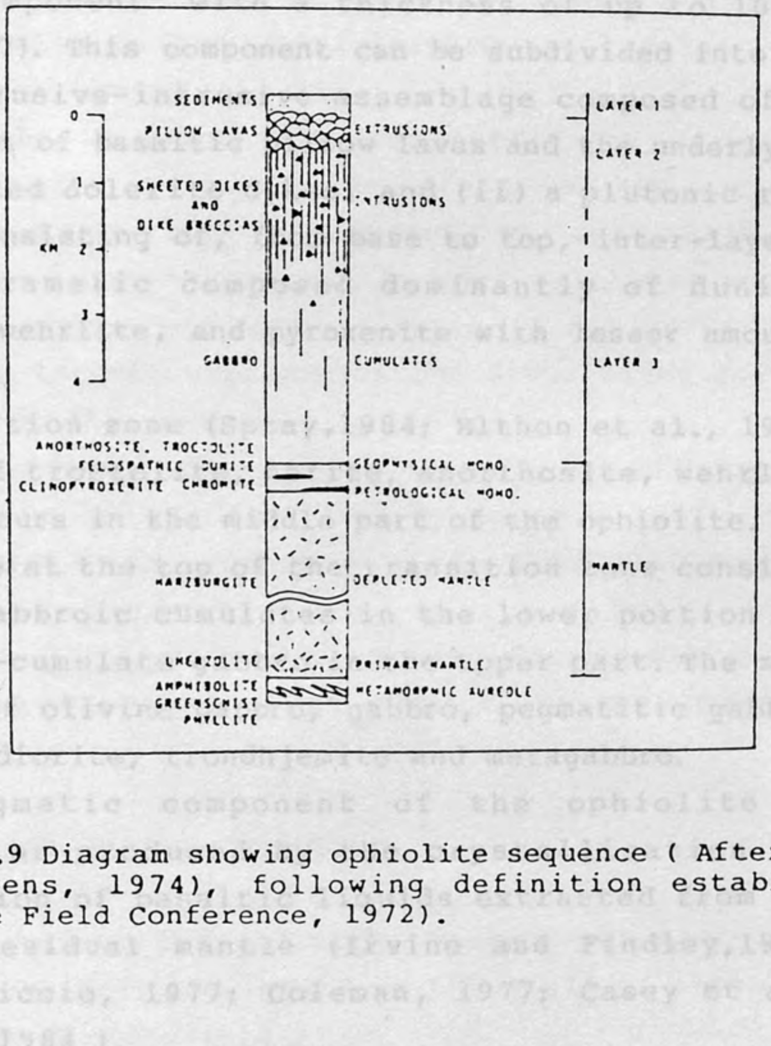


Fig. 3.9 Diagram showing ophiolite sequence (After Williams & Stevens, 1974), following definition established in Penrose Field Conference, 1972).

and wehrlite which are all invariably serpentinized. This component is interpreted to represent the residual mantle produced by partial fusion and the extraction of basaltic liquids (Irvine and Findley, 1972; Coleman, 1977). Most of the ultramafic rocks are massive and non-cumulate peridotites.

The upper part of the ophiolite consists of a "magmatic component" with a thickness of up to 10 km (Coleman, 1977). This component can be subdivided into (i) an upper extrusive-intrusive assemblage composed of an uppermost unit of basaltic pillow lavas and the underlying unit of sheeted dolerite dykes, and (ii) a plutonic rock assemblage consisting of, from base to top, inter-layered cumulate ultramafic composed dominantly of dunite, harzburgite, wehrlite, and pyroxenite with lesser amounts of websterite.

A transition zone (Spray,1984; Elthon et al., 1984) consisting of troctolite, norite, anorthosite, wehrlite and dunite occurs in the middle part of the ophiolite. The gabbroic zone at the top of the transition zone consists of layered gabbroic cumulates in the lower portion and massive, non-cumulate gabbro in the upper part. The zone is composed of olivine gabbro, gabbro, pegmatitic gabbro, anorthosite, diorite, trondhjemite and metagabbro.

The magmatic component of the ophiolite is interpreted as produced by the crystallisation and differentiation of basaltic liquids extracted from the underlying residual mantle (Irvine and Findley,1972; Church and Riccio, 1977; Coleman, 1977; Casey et al., 1981; Spray, 1984).

Rock association of the ophiolite in the East Arm of Sulawesi corresponds to ocean floor and has been formed at the oceanic ridge.

In addition to these assemblages, in many places, the ophiolite belt also contains,locally, amphibolites, metabasic rocks and metachert, such as those occurring in

Tanjung Api, Bunta River, and Siuna River just to the east of Siuna village (Fig.2.2). The metamorphic rocks occur as slivers and always in fault contact with ophiolitic rocks. They are well-foliated, sometimes isoclinally folded.

The amphibolites are dark in thin section with nematoblastic and xenoblastic textures, fine to medium grained, consisting essentially of hornblende and subsidiary feldspar. The hornblende shows a linear preferred orientation and the feldspars are tabular and multiply twinned grains.

The metabasic rocks are foliated; the original pyroxene, replaced by hornblende, preserves the ophitic texture and epidote-albite pseudomorphs replaced the basic plagioclase; chlorite is present in most metabasic rocks. The thickness of the metamorphic rocks is not definitely known, but the largest exposure in the Siuna River may be up to 200 m wide (Rusmana et al., 1984).

This metamorphic assemblage seems to be similar to the metamorphic soles described by Spray (1984) which forms a coherent basal sequence to the overlying mantle peridotites of the ophiolite. Alternatively, this metamorphic assemblage may have been formed along the transform faults in the ocean (Dr.A.J. Barber, person. comm.).

Emplacement of the Ophiolite

Numerous models for ophiolite emplacement have been proposed and put forward, which on the basis of tectonic process, can be divided into :

(i) Collision-subduction-obduction-accretion processes (Dewey and Bird, 1970; Church and Stevens, 1971; Davies, 1971; Oxburgh, 1972; Ernst, 1973; Christensen and Salisbury, 1975; Dewey, 1976; Smith and Woodcock, 1976; Gealey, 1977; Malpas and Stevens, 1977; Welland and

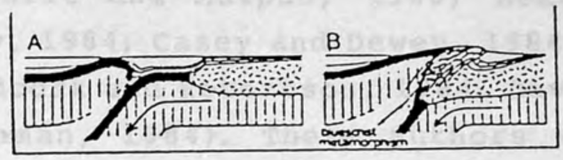


Fig. 3.10 Emplacement of ophiolite during thrusting oceanic crust and mantle onto continental crust (After, Dewey & Bird, 1970) A:Tethyan and B:Cordilleran type subduction zone.

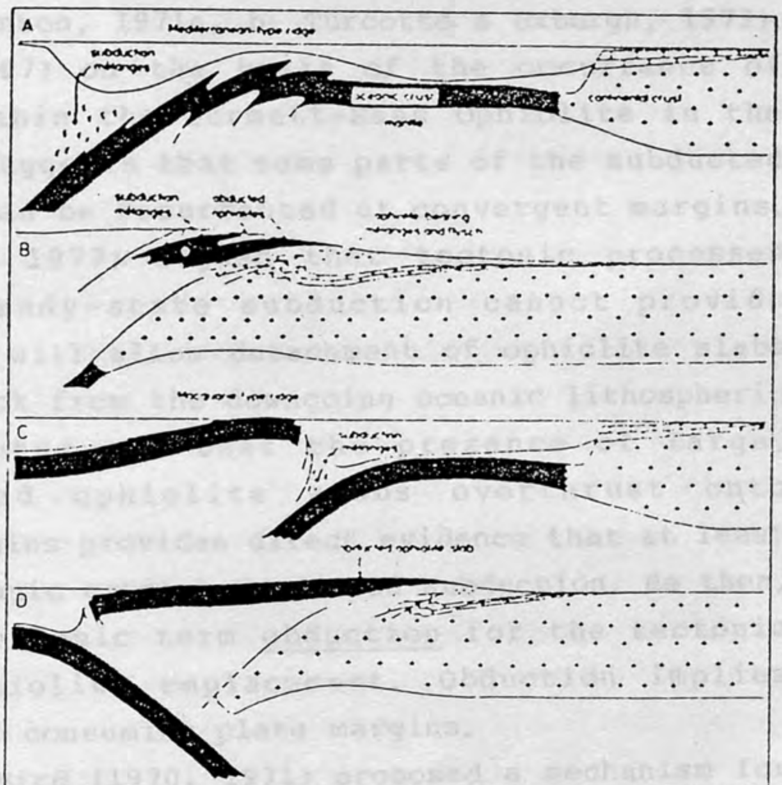


Fig. 3.11 Diagrams showing possible mechanism for the obduction of ophiolite onto continental margins (After, Dewey & Bird, 1971). A-C : Tethyan type, D : Cordilleran type subduction zone.

Mitchell, 1977; Searle and Malpas, 1980; Searle and Stevens, 1984; Spray, 1984; Casey and Dewey, 1984; Ogawa and Naka, 1984; Woodcock and Robertson, 1984; Davies and Jacques, 1984; Coleman, 1984). These authors are all concerned with two aspects of subduction polarity, in which the polarity may dips away from (Tethyan type) or towards the continent (Cordilleran type).

A steady-state subduction is required to consume the large amounts of oceanic crust developed at the spreading ridge. The consumption is visualised as a bending of the oceanic plate downward and its sinking into mantle, where it is assimilated and incorporated into mantle (Oliver et al., 1969; Dickinson, 1971a, b; Turcotte & Oxburgh, 1972).

Bearth (1967) on the basis of the occurrence of blueschists within the Zermatt-Saas Ophiolite in the western Alps, suggests that some parts of the subducted oceanic crust can be resurrected at convergent margins. Coleman (1971, 1977) argued that tectonic processes related to steady-state subduction cannot provide conditions that will allow detachment of ophiolite slabs up to 12 km thick from the downgoing oceanic lithospheric plate. He pointed out that the presence of large, unmetamorphosed ophiolite slabs overthrust onto continental margins provides direct evidence that at least some of the oceanic crust had escaped subduction. He then, introduced a tectonic term obduction for the tectonic process of ophiolite emplacement. Obduction implies overthrusting at consuming plate margins.

Dewey and Bird (1970, 1971) proposed a mechanism for thrusting oceanic crust and mantle onto a continental margin (Fig. 3.10, 3.11). Dewey and Bird (1971) and Coleman (1977) pointed out that there are several tectonic situations that would allow the detachment of oceanic crust prior to overthrusting (Fig. 3.11).

It is generally considered that the thickness of the oceanic plates ranges from 60 - 100 km (Oxburgh, 1974).

The thickest (i.e. 12 km) known obducted ophiolite slab is in Papua New Guinea (Davies, 1971). The emplacement of such thin ophiolite slabs requires some sort of detachment surface to develop within the top portion of the oceanic plate. Armstrong and Dick (1974) suggested that a steep geothermal gradient underlying the detached slab is necessary and described the detachment as follows : 'At the time of detachment, the relatively brittle and rigid cover moves away from its thermally softened base. At first, displacement between cover and base is penetrative, but as strain softening (and shear heating ?) proceeds, displacement becomes localised along a fault zone that will be parallel to isotherms in the rocks, and the overthrust crystalline-based sheet is thus freed from its base. During movement, the underlying rocks of the detached sheet may be metamorphosed, and fragments of the overridden rocks may be picked up'.

The detachment and obduction of young hot oceanic crust is suggested by Christensen and Salisbury (1975) as follows : ' The youth of the ophiolite imposes a considerable restraint on possible mechanisms of emplacement. During the closure of any ocean basin through subduction of one or both of its limbs, the ridge crest itself must at some point be subducted (Fig. 3.12). This point is unique in that for the first and only time a thin hot mechanically weak segment af oceanic crust and upper mantle, laced with magma chambers, is presented to the subduction mechanism. That subduction of the ridge crest occurs without incident is unlikely. It is anticipated, rather, the ridge crest will be dismembered by faulting, major segments, particularly from the upper levels of the outboard plate being obducted onto continental margins, while the inboard plate is depressed under the approaching continental plate and subducted'.

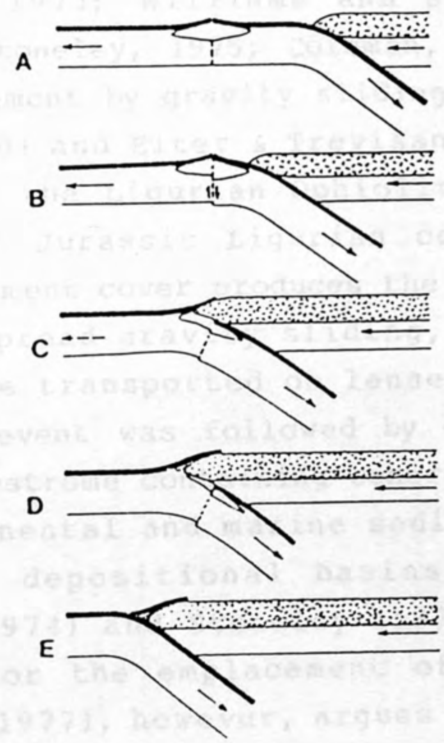


Fig. 3.12 Diagrams showing ophiolite emplacement during subduction of a ridge crest (After Christensen & Salisbury, 1975). A, B = Cordilleran, C, D = Tethyan, E : a combination of Tethyan and Cordilleran types subduction zone.

(ii) Gravity-sliding process (e.g. Abbate et al., 1970; Elter and Trevison, 1973; Williams and Smyth, 1973; Glennie et al.,1974; Stoneley, 1975; Coleman, 1977).

Ophiolite emplacement by gravity sliding is proposed by Abbate et al. (1970) and Elter & Trevisan (1973) with special reference to the Ligurian Ophiolite, northern Apennines. Uplift of Jurassic Ligurian oceanic crust together with its sediment cover produces the Bracco ridge giving rise to widespread gravity sliding, where large ophiolite blocks were transported on lenses of breccia (olistostrome). This event was followed by overthrusting of the Ligurian olistostrome containing coherent blocks of ophiolite over continental and marine sediments of the Tuscan and Umbrian depositional basins during the Tertiary. Glennie (1974) and Stoneley (1975) suggest a similar mechanism for the emplacement of the Semail Ophiolite. Coleman (1977), however, argues that neither explanation provides a satisfactory tectonic mechanism for elevating the ancient oceanic crust high enough to initiate gravity sliding.

(iii) Strike-slip ophiolite emplacement (e.g. Brookfield, 1977; Saleeby, 1977; Karson and Dewey, 1978; Robertson and Woodcock, 1980, 1981; Woodcock and Robertson, 1981, 1982, 1984; Casey and Dewey, 1984).

Evidence for strike-slip ophiolite emplacement in the Antalya Complex, SW Turkey has been described in detail by Woodcock and Robertson (1981, 1982) and Robertson and Woodcock (1980, 1981). Woodcock and Robertson (1984) describe a number of diagnostic features of strike-slip movement during ophiolite emplacement (Fig. 3.13):

1. An unexposed tectonic base to the ophiolite sheet is more likely in a strike-slip setting.
2. Syn-emplacement metamorphic rocks will usually occur along steep rather than shallow dipping fault zones in a strike-slip belt.

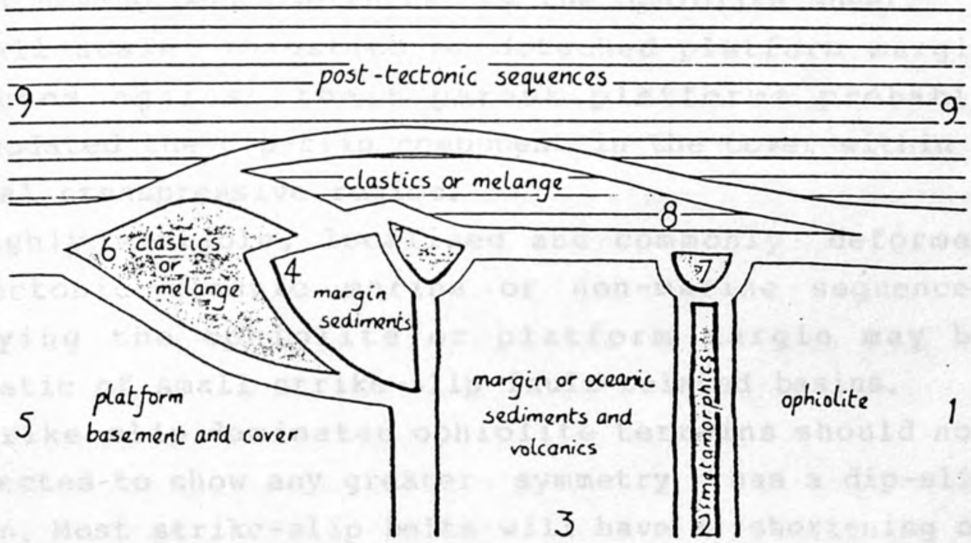


Fig. 3.13 Diagramatic exploded section across an ophiolite terrain dominated by strike-slip emplacement tectonics (After Woodcock & Robertson, 1984).

3. All or many of the pre-emplacement terrains or the syn- emplacement sequences may show steep faults or shear zones with strike-slip displacement.
4. Platform and platform margin terrains emplaced along with ophiolites by strike-slip tectonics may show no sign of having been overthrust by the ophiolite sheet.
5. Small-scale thrusting of detached platform margin sequences against their parent platforms probably accommodated the dip-slip component in the cover within a regional transpressive regime.
6. Highly variable, localised and commonly deformed syn-tectonic clastic marine or non-marine sequences overlying the ophiolite or platform margin may be diagnostic of small strike-slip fault-related basins.
7. Strike-slip dominated ophiolite terrains should not be expected to show any greater symmetry than a dip-slip terrain. Most strike-slip belts will have a shortening or extension component across them which will produce an asymmetric structure, and, failing that, late isostatic adjustments will usually modify any symmetric structure.

As discussed later, the last two processes are considered not to be the main mechanisms for the emplacement of the ophiolite belt in the East Arm of Sulawesi. These processes, however, may have been active independently or contemporaneously in some parts of the belt or at some time during the history of ophiolite emplacement.

In the following section the structural relationship of the ophiolite with the adjacent rock units within the collision zone is analysed and discussed in the light of the collision-subduction-obduction-accretion hypothesis, which is considered to be the main tectonic process for emplacement of the ophiolite belt.

Primary requirements for palinspastic reconstruction and study of processes leading up to ophiolite emplacement are an accurate and detailed geological map, good and updated biostratigraphic control of sediments both beneath and above the ophiolite, the construction of well-balanced cross-section, dating of ophiolite formation and associated rocks and seismic and gravity data for subsurface control.

The present detailed geological mapping on selected sections in the East Arm of Sulawesi provided more complete data and tighter structural control on a complete cross-section of the region. Unfortunately, so far, geophysical data of the East Arm are not available, but gravity data of Silver et al. (1978) crossing Central Sulawesi from west to east and seismic data of McCaffrey et al.(1982) from Tomini Gulf and the North Arm of Sulawesi are used to give a better insight into the subsurface structure of the ophiolite belt.

Along the Balantak coast the ophiolite is juxtaposed with imbricated Jurassic to Palaeogene continental margin carbonate-dominated sequences (i.e. Sinsidik Beds, Luok Beds and Salodik Limestones), and is unconformably overlain by the Late Miocene to Pliocene volcanogenic Lonsuit Turbidites. The ophiolite and the carbonate sequences are highly deformed, repeatedly faulted and the carbonates are tightly to recumbently folded and thrust, which gives rise to the highly imbricated nature and the duplex structure of the sequences. By contrast, the Neogene sediments are much less-deformed and only gently buckled.

A narrow zone of melanges occurs along the contact of the ophiolite and the continental margin sediments. The contact zone is marked by the Balantak Fault System, which appears to have been formed originally as a thrust fault (i.e. as part of the Batui Thrust), but since Plio-Pleistocene time it has become a dextral strike-slip fault. This fault seems to be also responsible for the occurrence of only a narrow zone of ultramafic rocks in Poh Head, the main part of which now occurs, at least 50 km further to the west.

Tectonostratigraphically, 4 distinctive tectonic units occur in the Balantak area (Fig.3.4; 3.16):

1. Cenomanian to Eocene upper part of the ophiolite (gabbroic-sheeted dolerite dyke-pillow basalt).
2. Late Jurassic to Palaeogene continental margin carbonate-dominated sequences.
3. Melange which is considered to be similar to that occurring further to the west (i.e. Kolokolo Melange of late Middle Miocene to Pliocene age), and
4. Late Miocene to Pliocene volcanogenic Lonsuit Turbidites.

Structures and nature of the exposures, biostratigraphy and tectonostratigraphy of rock units and a well-balanced cross-section along the Balantak coast from north to south shown in Fig. 3.4 and 3.17 suggest that the first stage of tectonic process was essentially collision of the ophiolite belt against a continental margin. The collision is marked and documented by the occurrence of melange, which has been formed since late Middle Miocene time. The collision episode was then followed subsequently by upthrust movement of the ophiolite onto the continental margin sequences (i.e. obduction process), giving rise to the nappe-like structure across the Balantak region (Fig.3.16). The obduction episode may have been followed by subduction and accretion processes, which will be discussed together with the other sections studied in detail at the end of this chapter.

In Biak-Poh section (Fig.3.5), four tectonostratigraphic units can be readily recognised : 1. ophiolites dominated by gabbroic rocks and to the east (i.e. Siuna river) the ophiolite belt contains serpentinized peridotites, some 300 m thick, which are associated with an imbricated metamorphic sole. 2. Palaeogene Salodik Limestones and Late Cretaceous Luok

Beds which occur in narrow, fault bounded exposures, representing the continental margin carbonate sequences.
3. melange, some of which has a red scaly clay matrix, and
4. Late Miocene to Pliocene molasse-type sediments, Biak Conglomerates, which unconformably overlie Salodik Limestones.

The ophiolite and the carbonate sequences are highly deformed, faulted and repeatedly thrust. The highly imbricated nature of the region is clearly shown in the aerial-photographs in Fig. 3.15 and Plate 3.14 is discussed in more detailed in another section of this chapter.

As in the Balantak Section, the structures and nature of different rock units which are juxtaposed in the Biak-Poh area, combined with biostratigraphy and tectonostratigraphy of these rocks, all suggest that the first tectonic episode was due to collision of the ophiolite belt against the continental margin.

Failure and sudden collapse of a previously stable continental margin, subsequent to collision with the ophiolite belt is documented by the occurrence of spectacular carbonate slope and platform deposits (i.e. Nambo and Luok Beds and Salodik Limestones) along the southern margin of the Batui Thrust-Balantak Fault System, forming most of the mountain ranges within half the southern portion of the East Arm. The fact that the melange contains clasts detached from both the ophiolite suite and the continental margin sequences along the contact zone, strongly suggests that the Cretaceous-Palaeogene shelf edge collided with the subduction complex, and that this collision was responsible for the original emplacement of the ophiolites.

In the Kolo Atas area, the ophiolite is juxtaposed with the Palaeogene Salodik Limestones, and the quartz-dominated sediments of Jurassic Kapali Beds, and further to the west, the Triassic Lemo Beds (i.e. part of the

Tokala Formation, Simandjuntak et al.,1983). The contact zone is always associated with the Kolokolo Melange with a matrix of lime-mudstone and marlstone rich in planktonic foraminifera of late Middle Miocene to Pliocene age, and clasts detached from the ophiolite belt and continental margin sediments.

The tectonostratigraphy of Kolo Atas extended to the Tokala area can be divided into: 1. highly deformed, tightly folded and faulted Cretaceous radiolarian chert and calcilutite which represent deep sea sediments deposited on top of the ophiolite suite. 2. Triassic to Palaeogene continental margin sequences. 3. ophiolite and its associated metamorphic rocks. 4. late Middle Miocene-Pliocene Kolokolo Melange, and 5. Late Miocene to Pliocene molasse-type sediments (i.e. Kolo Beds).

As in Poh Head, these rock units are all juxtaposed in the Kolo Atas and the Tokala areas. Structures, biostratigraphy, and tectonostratigraphy strongly suggest that the first tectonic episode was collision of the ophiolite suite against the Banggai-Sula Platform. The collision is documented by the occurrence of Kolokolo Melange which has been formed since late Middle Miocene. The age of the fossils in the melange matrix also suggest that collision might have been continuously active until Pliocene time.

Data obtained from these detailed sections described above suggest that the first stage of collision in Middle Miocene is a Tethyan-type collision, in which the ophiolite belt (i.e. ESOB) overthrust onto the Banggai-Sula Platform. The Eastern Sulawesi Ophiolite Belt (ESOB), therefore, can be classified as obducted ophiolite suite.

Generally, continental crust is thought to be too buoyant to be subducted beneath oceanic crust (Condie, 1982), and hence when continental crust arrives at a subduction zone, a collision takes place. It appears that the original emplacement of the ophiolite occurred during this collision.

As the collision proceeded, the obduction process took place, in which the ophiolite suite was thrust onto the continental margin. This tectonic episode led to the original emplacement of the ophiolite suite. The obduction process has led to the imbrication of both the ophiolite belt and the continental margin sequence. Molnar and Gray (1979) pointed out that due to buoyancy and temperature constraints, the underplating continental crust may extend up to only several hundreds of kilometres. Seismic data (McCaffrey et al., 1982), however, suggest that beneath the ophiolite the collision in the East Arm of Sulawesi may be still active at the present time. This will be discussed further in Chapter 5.

Emplacement of an ophiolite due to continental underthrusting of an accretionary prism occurs in many parts of the world. Such as that the Northern Australian margin underplating Timor and the Banda Arc (Audley-Charles et al.,1972; Carter et al., 1976; Barber et al., 1977), the South China margin underthrusting the Taiwan arc (Karig, 1973; Roeder, 1977; Bowin et al., 1978), and the Papua New Guinea ophiolites emplaced by underthrusting of the Northern Australian margin beneath an Eocene island arc (Davies and Smith, 1971; Hamilton, 1979). In all these cases, a continent has been subducted beneath the oceanic floor.

Large allochthonous ophiolite nappes obducted onto continental margins occur in the Bay of Islands, Newfoundland, Canada (Casey and Dewey, 1984), Semail Ophiolite in Oman (Coleman, 1981) and Spontang Ophiolite, Western Himalaya (Fuchs, 1979; Searle,1983).

3.4 STRUCTURES PRODUCED BY CONVERGENCE IN THE EAST ARM.

3.4.1 INTRODUCTION

Structural and tectonic setting of the East Arm of Sulawesi show a typical collision complex, which consists of two distinctive structural domains, namely, (i) imbricated complex with an allochthonous Triassic to Palaeogene continental margin sediments (Balantak Group) juxtaposed with ophiolite belt, and (ii) autochthonous Neogene coarse clastic sediments and volcanogenic turbidites (Batui Group).

As described in Chapter 2, the continental margin sediments, including Lemo Beds, Kapali Beds, Sinsidik Beds, Nambo Beds, Luok Beds and Salodik Limestones all occur in fault slivers or fault-bounded exposures and form the imbricated complex in the East Arm of Sulawesi. Folding, fault and thrust of all scales occurred repeatedly within the region. Mesoscopic structures including joints and fractures occur in all rock units, while cleavage, shearing or foliation occur only locally.

In the following section structures of the imbricated complex are discussed, while the structures of the Neogene coarse clastic rocks will be discussed in Chapter 4.

3.4.2 FOLDING

Most of the sedimentary rocks are well-bedded, but bedding attitudes change abruptly over very short distances. Most of the sediments are generally moderately to steeply dipping towards northwest, north and north-northeast, and fold axes are moderately to steeply plunging towards southwest, west or east, southeast direction.

The outcrops show that the folding is mostly on a mesoscopic scale, while major folding (regional fold)

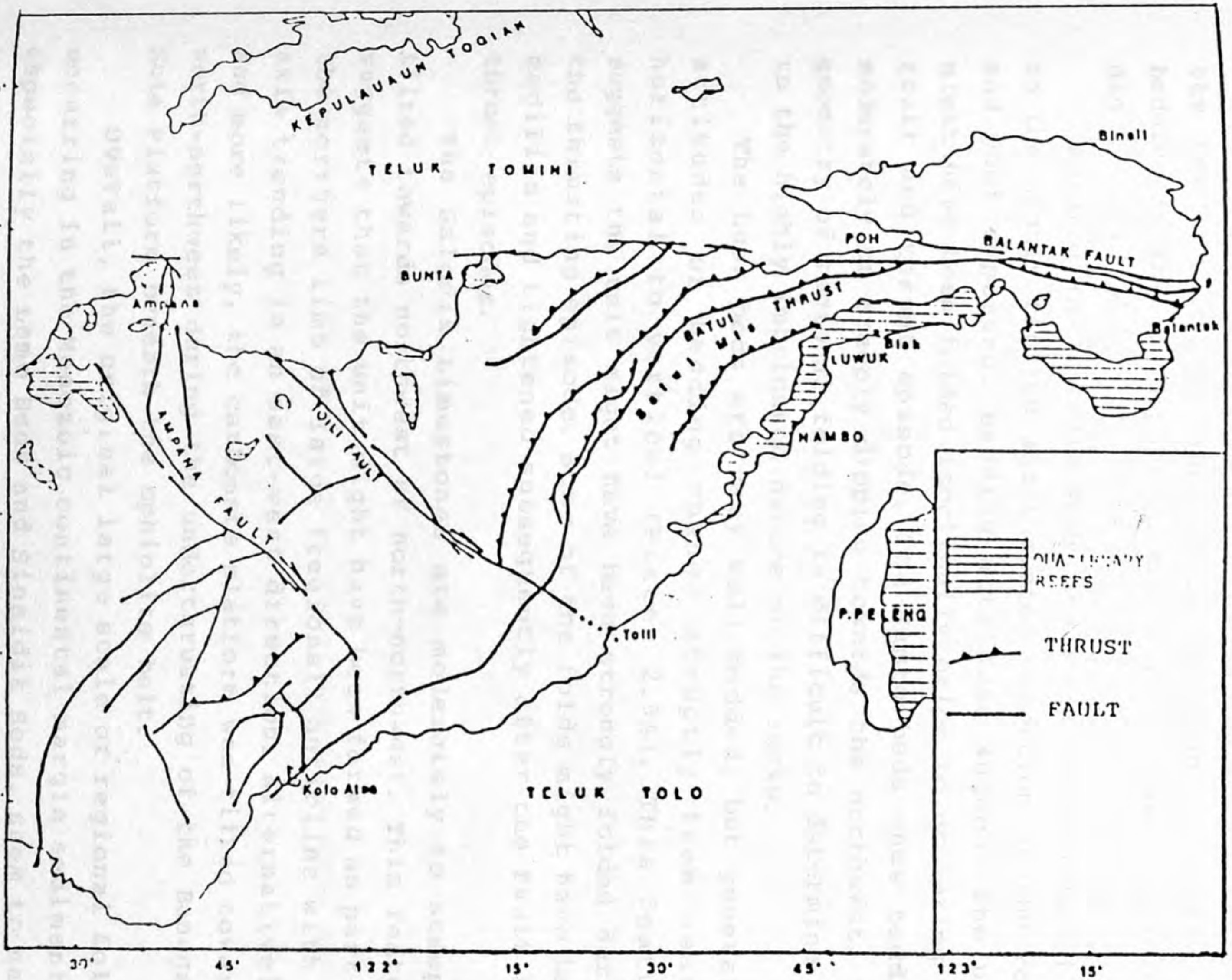


Fig. 3.14 Map showing structural configuration of the East Arm

cannot be determined due to the highly imbricated nature of the sediments, especially the Mesozoic to Palaeogene sequences.

No regional large scale folding can be recognised in the Lemo Beds; the succession is highly faulted and bedding invariably dips towards north-northwest. Some beds dip vertically.

Folding in Sinsidik Beds could not be deciphered due to the highly faulted and thrusted condition of the rocks and poor exposure. Bedding attitudes suggest the unit might have been folded isoclinally prior to or during the fault and thrust episode. The Nambo beds show bedding moderately or steeply dipping towards the northwest. The geometry of regional folding is difficult to determine due to the highly imbricated nature of the rocks.

The Luok beds are very well-bedded, but generally attitudes of bedding change abruptly from nearly horizontal to vertical (Plate 2.8A). This feature suggests the unit might have been strongly folded during the thrusting episode. Some of the folds might have been modified and tightened subsequently after the fault and thrust episodes.

The Salodik Limestones are moderately to steeply tilted towards northwest or north-northwest. This feature suggests that the unit might have been formed as part of the northern limb of large (regional) anticline with an axis trending in an east-west direction. Alternatively, and more likely, the carbonate platform was tilted towards north-northwest during the underthrusting of the Banggai-Sula Platform beneath the ophiolite belt.

Overall, the original large scale or regional folds occurring in the Mesozoic continental margin sediments, especially the Lemo Beds and Sinsidik Beds, seem to have been greatly disrupted by subsequent fault and thrust. Hence, their present fault-sliver exposures would not show the original attitude and geometry of the regional folds.

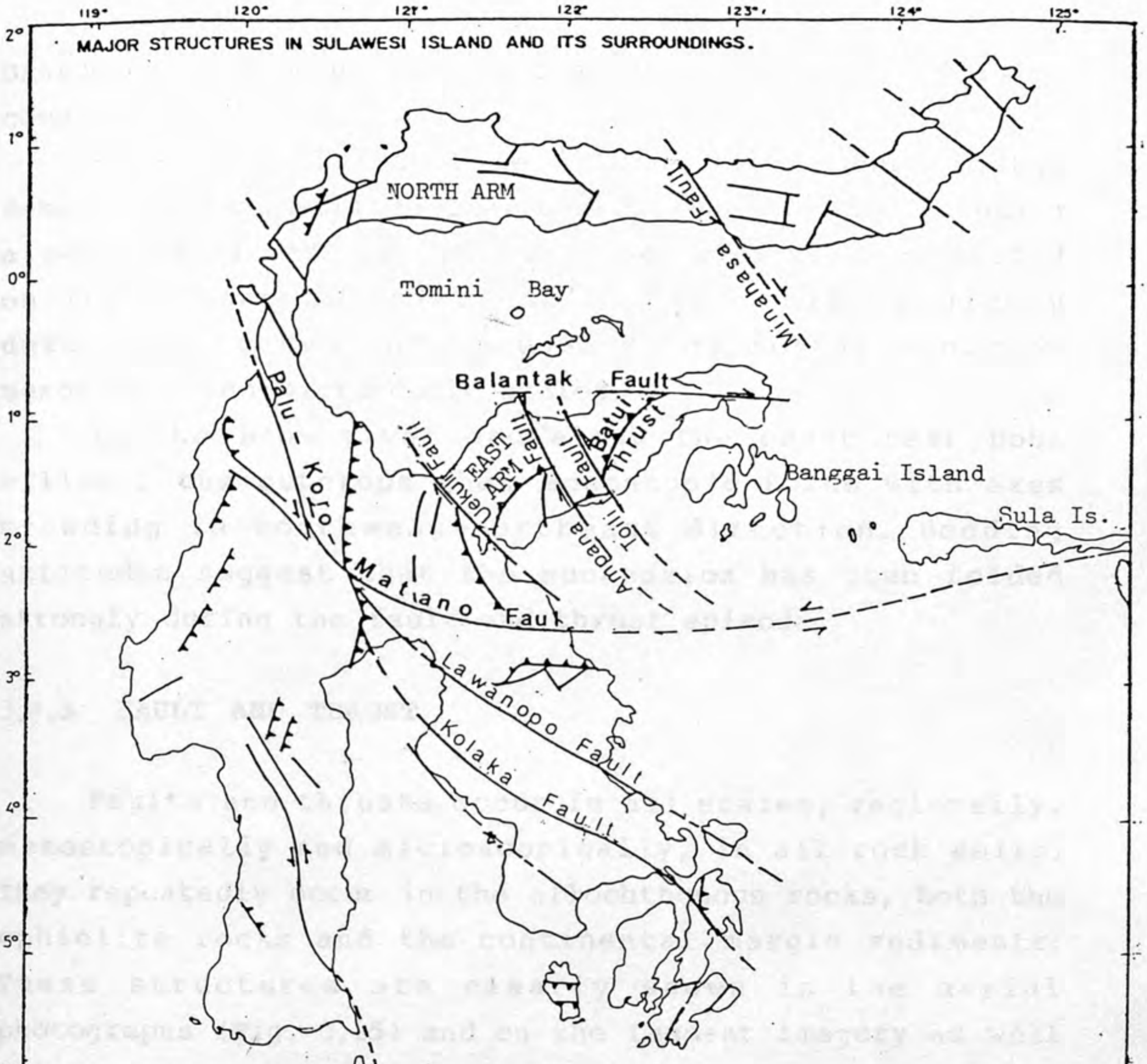


Fig. 3.14.2 Map showing major structures in Sulawesi and its surroundings.

Disrupted fold structures is characteristic of a collision complex.

Similarly, large scale folding is not seen in the Boba Beds. As described previously, the Boba Beds contain a well-bedded radiolarian chert and calcilutite deposited on top of the ophiolite suite. The unit is highly deformed; folded and thrust. Folds mostly occur on mesoscopic and microscopic scales.

In the Boba river and along the coast near Boba village, the outcrops show mesoscopic folds with axes trending in southwest-northeast direction. Bedding attitudes suggest that the succession has been folded strongly during the fault and thrust episode.

3.4.3 FAULT AND THRUST

Faults and thrusts occur in all scales, regionally, mesoscopically and microscopically, in all rock units. They repeatedly occur in the allochthonous rocks, both the ophiolite rocks and the continental margin sediments. These structures are clearly shown in the aerial photographs (Fig. 3.15) and on the Landsat imagery as well (Plate 3.14).

In outcrops, fault zones are usually marked by the occurrence of fault breccia and gouge, mylonites, slickensides, abruptly changing bedding attitudes over very short distances, and the outcrop of fault (or thrust) planes are commonly marked by steep cliffs. For the purposes of description and interpretation the major (regional) thrust occurring in the East Arm of Sulawesi are named **Batui Thrust.**

BATUI THRUST

The Batui Thrust is a long, arcuate and convex northwest thrust fault. The south-end of the thrust dies

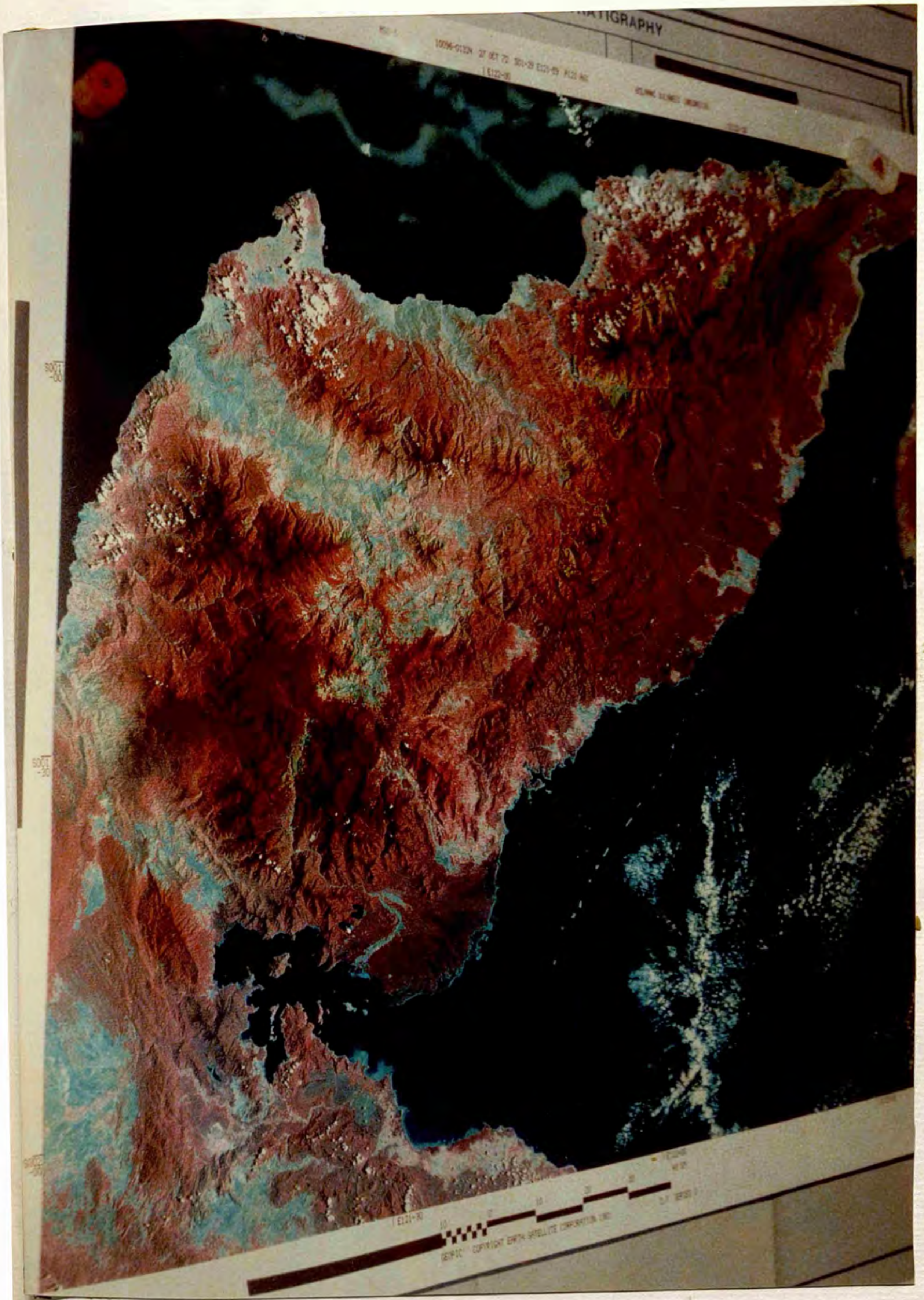


Plate 3.14 Landsat imagery showing the imbricate nature of the East Arm of Sulawesi.

out in Tolo Bay, while in the Poh Neck it merges with the Balantak Fault System, which was considered originally to be part of the Batui Thrust. The Batui Thrust stretches to the east, off-shore into the North Banda Sea and terminates at the East Sangihe Fault (Hamilton, 1973; Silver et al., 1983), and is up to 100 km long.

The Batui Thrust - Balantak Fault System marks the zone of convergence between the Banggai-Sula Platform and the Balantak Ophiolite Belt. The zone ranges in width from several hundred metres on the Balantak coast to more than 1 km in the Toili area (Fig. 3.14). On the regional scale the fault zone is shown as a single line, but in exposures it consists of several faults and thrusts, all dipping towards north and northwest. In this study as discussed later in Chapter 4, the eastern portion of the thrust (i.e. the Balantak Fault System) on the basis of field observation, operated as a dextral strike slip fault, at least during the later stages of its movement.

The fault zone is associated with the Kolokolo Melange. In many places, the matrix of the melange is sheared and some of the clasts are boudinaged. In the Kolo Atas area, the dip of the foliation suggests that the thrust plane dips moderately towards northwest . In Poh Neck and Balantak coast, the fault planes dip more steeply towards north. This probably, is due mainly to a change in fault movement from vertical to horizontal of the Balantak Fault System.

In the Tokala area, the ophiolite belt is thrust over the Triassic Tokala Formation. In Kolo Atas, Toili and Poh Neck, the exposures show that the upper part of the Palaeogene Salodik Limestones are underthrust beneath the Balantak Ophiolite. On the Balantak coast, the ophiolite is also thrust over the Late Jurassic Sinsidik Beds, Late Cretaceous Luok Beds and Palaeogene Salodik Limestones. These features, all suggest that the upthrust movement of the ophiolite belt onto the continental margin occurred in

Middle Miocene time, as the younger rocks involved within the thrust is Eocene to Early Miocene Salodik Limestones. This is also suggested by the occurrence of melange containing matrix of calcareous mudstone rich in planktonic foraminifera of late Middle Miocene to Pliocene age.

3.4.4 IMBRICATED COMPLEX

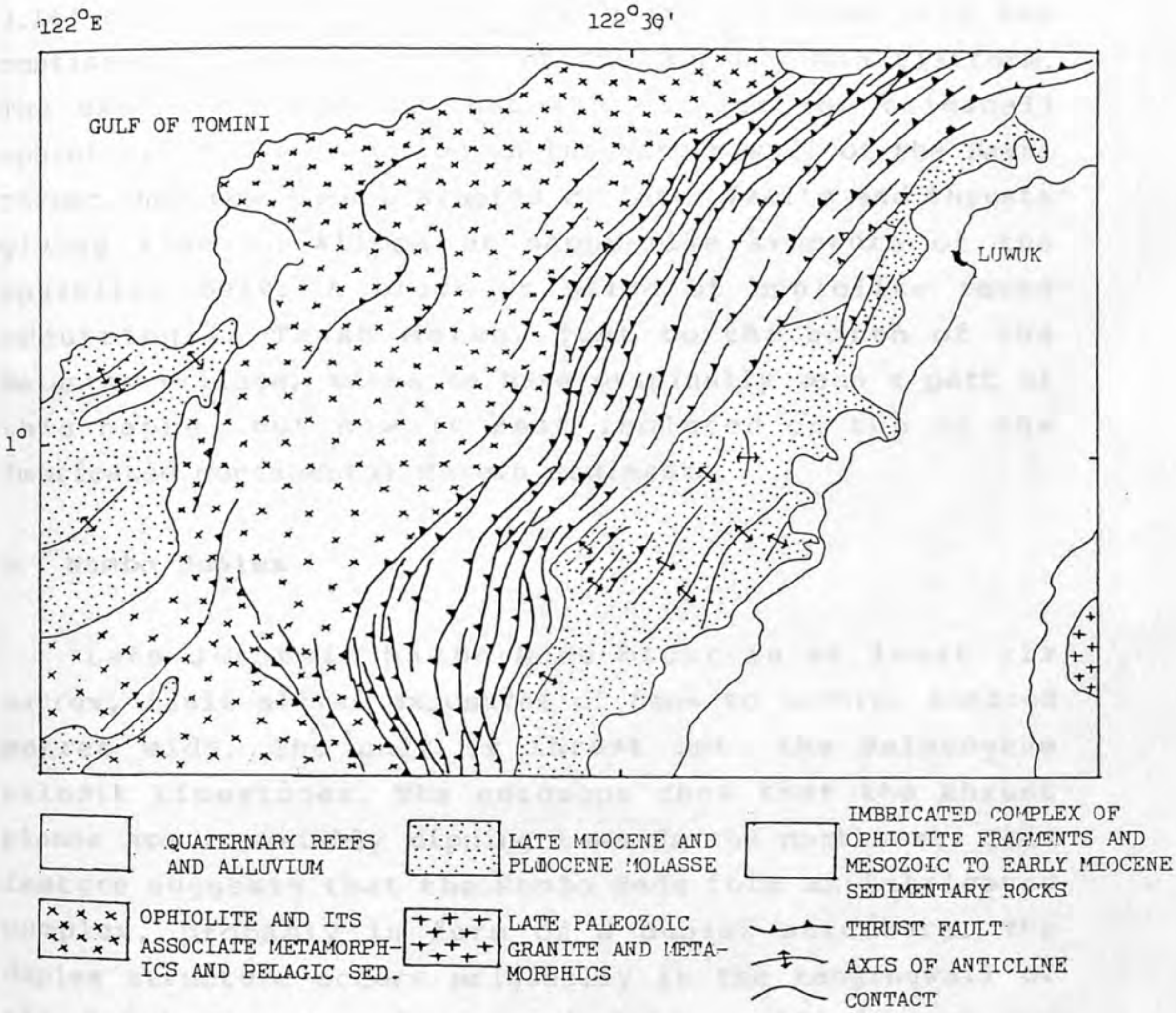
The imbricate structure is clearly seen in the outcrops, especially in the middle and the eastern part of the East Arm of Sulawesi. In particular areas, such as that in Nambo river and Balantak coast, the imbricate zone may show duplex structures, in which slices of continental margin sediments and ophiolitic rocks are thrust against each other. The imbricated complex is also clearly shown in the aerial photographs (Fig. 3.15) and on the Landsat imagenary (Plate 3.14).

In Nambo river, the exposures show the imbricate structure is marked by the occurrence of steep cliffs and water falls up to tens of metres high forming the exposure of thrust planes. The thrust planes are usually slickensided showing vertical movement. The thrust planes are dipping moderately to steeply towards northwest or north. The mesoscopic thrusts and faults have occurred during and subsequent to the major thrust episode.

A. Balantak Duplex

On the Balantak coast, the Late Jurassic Sinsidik Beds, Late Cretaceous Luok Beds, Palaeogene Salodik Limestones and Neogene Kolo Beds form an imbricated complex. The imbricated complex appears to have been developed in the hangingwall of the original major thrust (i.e. Batui Thrust). The exposures show that each rock unit is bounded by thrust to each other. The thrust

Fig. 3.15 Aerial-photograph interpretation map showing the imbricated complex in the East Arm of Sulawesi (modified after Inco, 1972; Hamilton, 1979; Surono et., 1984; Rusmana et al, 1984).



surfaces usually occurred in the form of footwall thrust. The exposures suggest that the whole thrust package were sliced and imbricated with thrust planes invariably dipping toward north or north-northwest.

A line-section drawn across the Balantak coast (Fig. 3.16) shows that the ophiolite belt is thrust over the continental margin sequence of the Banggai-Sula Platform. The exposures suggest that the earlier (or original) ophiolitic rocks occurring in the hangingwall of the Batui Thrust, has now been disrupted by later faults and thrusts giving rise to klippe or nappe-like sructure of the ophiolite belt. A block or slice of ophiolite rocks occurring in Tanah Merah, just to the north of the Balantak village, seems to have originally been a part of this nappe, but now it rest isolated on top of the imbricated continental margin sediments.

B. Nambo Duplex

Late Jurassic Nambo Beds occur in at least six narrow, fault-sliver exposures of tens to several hundred metres wide. The unit is thrust onto the Palaeogene Salodik Limestones. The outcrops show that the thrust planes are invariably dipping towards the northwest. This feature suggests that the Nambo Beds form an imbricated complex, probably in form of a duplex structure. The duplex structure occurs originally in the hangingwall of the Batui Thrust, which marked the collision of the Banggai-Sula Platform against the ophiolite belt.

The duplex contains the imbricated slices of Nambo Beds and Salodik Limestones in between a floor thrust and roof thrust. Formation of the Nambo Duplex is essentially related to the stacking of several thrust sheets of Nambo Beds and Salodik Limestones making up an imbricate zone. At the top of the duplex is bounded by the original major thrust surface of the Batui Thrust, but at the bottom, the

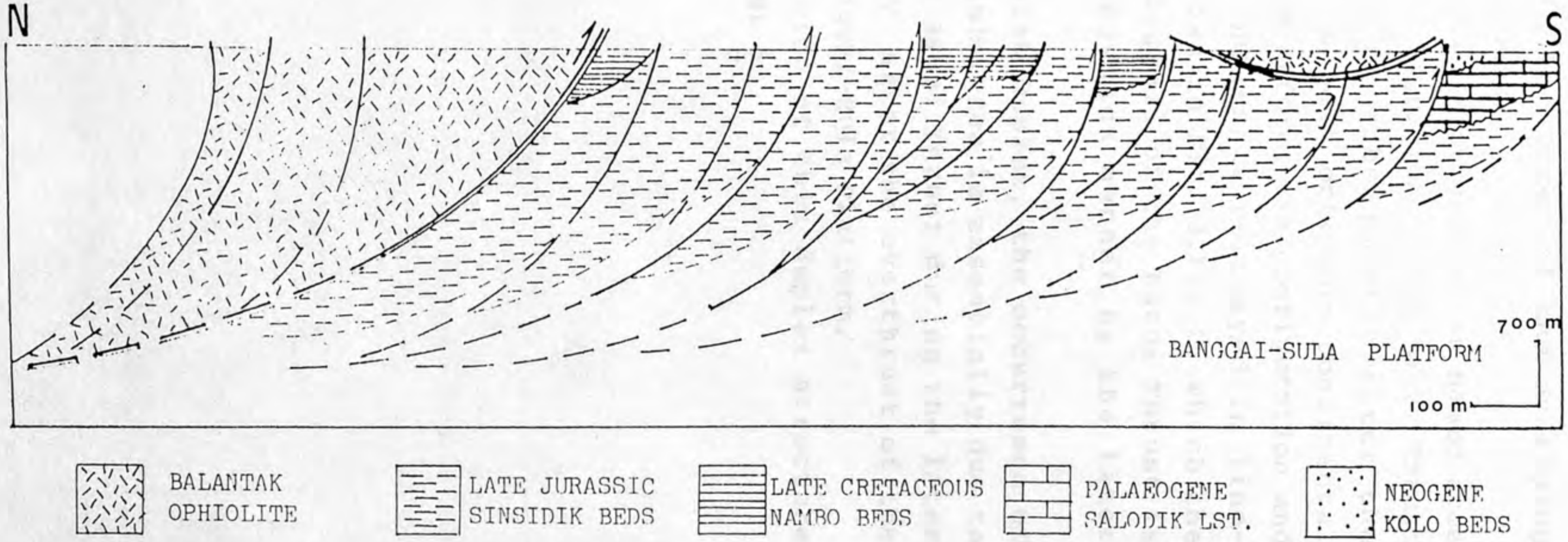


Fig. 3.16 Idealised cross-section across Tanjung Saro-Dua Islands, north of Balantak, showing the development Balantak Duplex, in which the continental margin sediments are inbricated and juxtaposed with the Balantak Ophiolite.

contact is not seen due to the cover of the overlying Neogene and Quaternary sediments.

The development of of duplex structure in Nambo area is mainly due to the forward migration of the Batui Thrust during the latter stages or subsequently after the overthrust of the ophiolite belt onto the continental margin of the Banggai-Sula Platform. The configuration and development of the Nambo Duplex is portrayed in line-section of Nambo river shown in Fig. 3.17, in which the rlier (or original) hangingwall of the Batui Thrust is carried forward in a piggyback manner by the later thrusts.

Similar to the Balantak Duplex, the occurrence of duplex structure in the Nambo area is essentially due to the forwards migration of Batui Thrust during the later stages and/or subsequently, after the overthrust of the ophiolite belt onto the Banggai-Sula Platform.

The tectonic implication of this duplex structure will be discussed in Chapter 5.

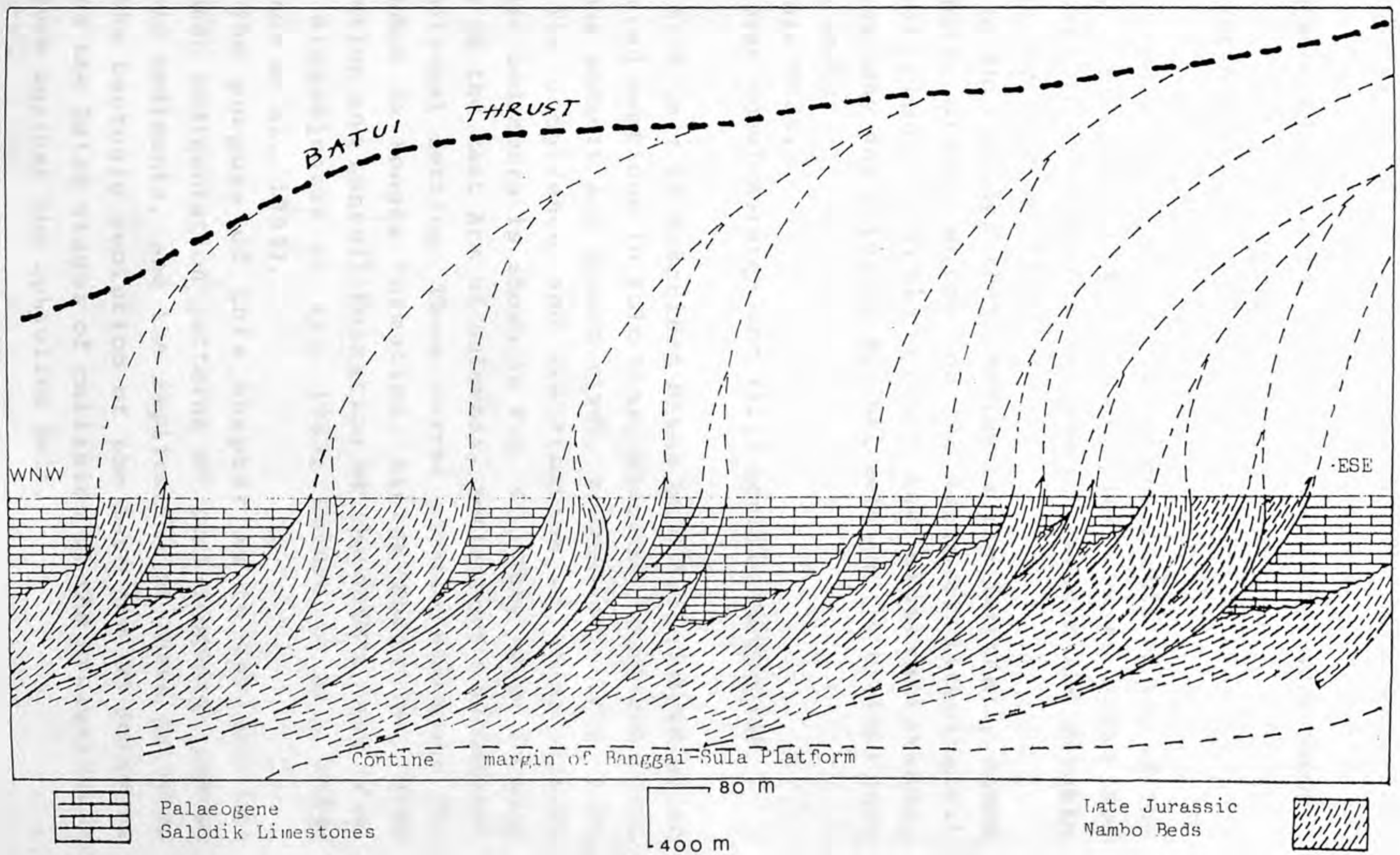


Fig. 3.17 Idealised cross-section of Nambo River, showing the imbricated complex of Late Jurassic and Palaeogene Salodik Limestones which formed the Nambo Duplex.

3.17

CHAPTER 4

SEDIMENTOLOGIC AND TECTONIC EVOLUTION OF THE BATUI GROUP

4.1. INTRODUCTION

Coarse clastic rocks rest unconformably on top of the eastern Sulawesi collision complex. These rocks are included in the Celebes Molasse of Sarasin and Sarasin (1901) and Kundig (1941).

In this study, these sediments are informally named the Batui Group, which on the basis of lithological association, sedimentology and biostratigraphy, physiography and inferred basinal setting, is divided into three units :
(i) Kolo Beds,
(ii) Biak Conglomerates and (iii) Lonsuit Turbidites.

Each unit is described based on detailed studies of measured sections in Kolo Atas, along the Biak-Poh road, Tanjung Lonsuit and Bombon river, respectively (Fig. 1.3).

The occurrence and distribution of these coarse clastic sediments is shown in Fig. 4.1. They occur in many parts of the East Arm of Sulawesi, reflecting a subbasinal depositional setting. These coarse clastic sediments are included in Bongka Formation, Kintom Formation, Biak Formation and Lonsuit Formation of the GRDC's maps (Fig. 4.2, Simandjuntak et al., 1983; Surono et al., 1984; Rusmana et al., 1984).

The purpose of this chapter is to analyse the regional sedimentation patterns of post-orogenic coarse clastic sediments, and the implication of these patterns for the tectonic evolution of the East Arm of Sulawesi during the later stages of collision of the Banggai-Sula Platform against the ophiolite belt.

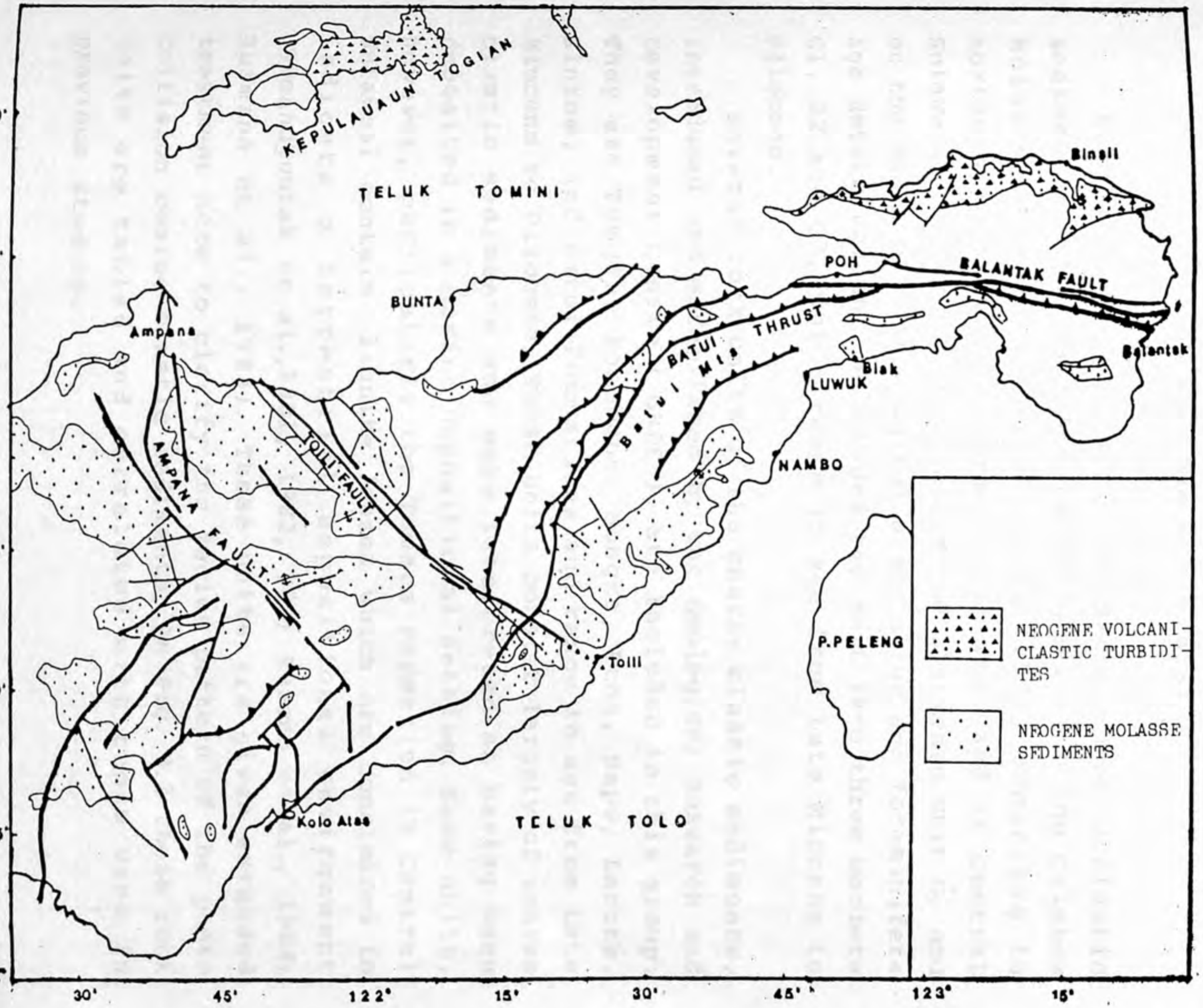


Fig. 4.1 Map showing the occurrence of Neogene post-orogenic coarse clastic rocks in the East Arm of Sulawesi (After Surono, 1981).

4.2. PREVIOUS STUDIES OF THE POST-COLLISION SEDIMENTS OF THE EAST ARM OF SULAWESI

Kundig (1941) correlated the Neogene coarse clastic sediments of the East Arm of Sulawesi with the Celebes Molasse of Sarasin and Sarasin (1901) occurring in northern part of South Arm of Sulawesi and in Central Sulawesi. Kundig (op cit) called the sediments Unit G, and on the basis of litthological association and foraminifera age determination, he divided the unit into three members Gl, G2 and G3, which range in age from Late Miocene to Pliocene.

Several rock units of the coarse clastic sediments, introduced and established by the Geological Research and Development Centre (GRDC), are included in this group. They are Tomata, Bonebone, Bongka, Puna, Napu, Larone, Kintom, and Batui Formations and range in age from Late Miocene to Pliocene. These units consist largely of coarse clastic sediments and were interpreted as having been deposited in a marine depositional setting. Some units, however, particularly the Tomata Formation in Central Sulawesi contain lignite lenses which are considered to indicate a terrestrial depositional environment (Simandjuntak et al.,1981, 1982, 983; Surono et al., 1984; Rusmana et al., 1984). These units are given extended treatment here to clarify the facies pattern of the post-collision coarse clastic sediments. In Fig. 4.2 these rock units are tabled and correlated with those used in previous studies.

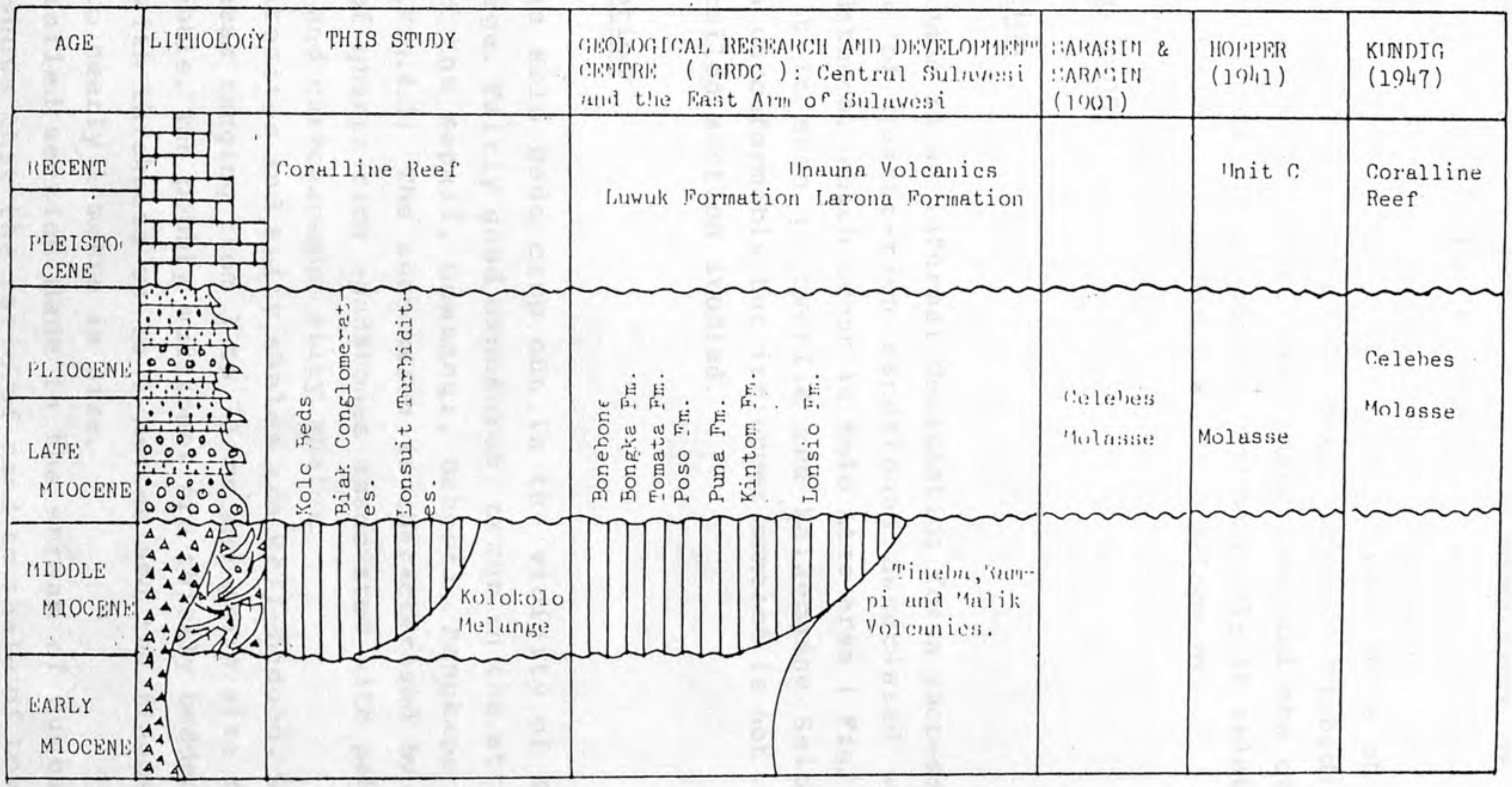


Fig.4.2 Stratigraphy of Neogene sediments in the East Arm of Sulawesi and their correlation with the previous work in the region.

4.3. STRATIGRAPHY, SEDIMENTOLOGY AND PETROLOGY OF THE BATUI GROUP

In the following discussion, three rock units of the Batui Group including the Kolo Beds, Lonsuit Turbidites and Biak Conglomerates are fully described and the other rock units are briefly discussed, particularly in relation to their facies patterns and basinal development.

4.3.1. KOLO BEDS

A. Definition

Kolo Beds is an informal designation for a succession consisting of quartz-rich sandstones associated with pebbly mudstones, which occur in Kolo Atas area (Fig. 4.3). The unit is seen to overlie the Palaeogene Salodik Limestones unconformably but its upper contact is not seen in the detailed section studied.

B. Description

The Kolo Beds crop out in the vicinity of Kolo Atas village. Fairly good exposures occur in the stream courses of the Kapali, Gombangi, Bahooti, Pangkape and Baholusa (Fig.4.3). The succession is characterised by the presence of quartz-rich sandstones associated with pebbly mudstones and carbonaceous silty shales.

The arenites and silty shales are well-bedded, with bed thickness ranging from 3 to 70 cm and grain size from silt to pebble. The pebbly mudstones are thickly bedded or massive, with thickness up to 6 metres and the fragments grade up to nearly 1 metre in size.

A detailed section made in the stream of Gumbangi (Fig.4.4) shows that the ratio of sand to shale of Lovell is high (4:1). The proximality index (Pl) of Walker (1967)

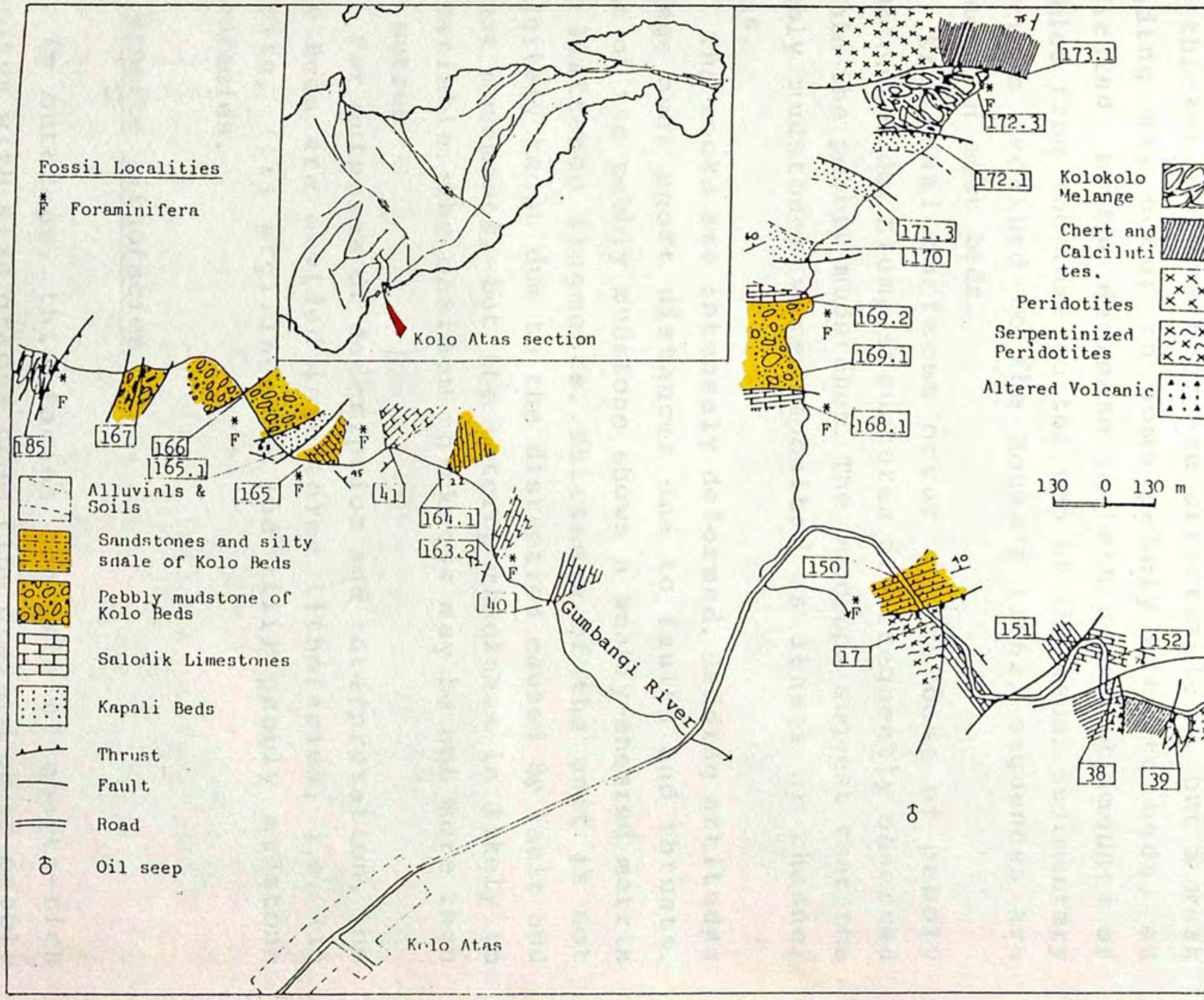


Fig. 4.3 Geological traverse map of Kolo Atas area, showing the occurrence of Kolo Beds.

of the succession is also very high, up to 85%.

Thin parallel laminae occur in some beds of arenite and silty shale and range in thickness from few a millimetres to 2 cm. Current ripple laminae are also present in some beds of arenite and silty shale. Most of the thicker arenite beds are structureless, but a weak grading may occur in some pebbly arenite beds, as indicated by the decrease in both size and amounts of pebbles from the base to the top of the beds. Sedimentary features ascribed to the Bouma's (1962) sequences are present in most beds.

Erosional surfaces occur at the base of pebbly mudstone beds. Slump structures are frequently observed within the pebbly mudstones. The outcrops suggest that the pebbly mudstones were deposited as lenses or channel fills.

The rocks are intensely deformed. Bedding attitudes change over short distances due to faults and thrusts. Some of the pebbly mudstone shows a weakly-sheared matrix and flattened fragments. Thickness of the unit is not definitely known due to the disruption caused by fault and thrust movements, but the outcrop thickness is likely to be variable. The maximum thickness may be not more than 300 metres.

For purposes of description and interpretation, the Kolo Beds are divided into three lithofacies, i.e. (i) arenite, (ii) argillaceous and (iii) pebbly mudstone lithofacies.

(i) Arenite lithofacies

In outcrops, this facies consists of quartz-rich arenites with size grades from fine grained up to pebble size in the lower part, which is gradually replaced by calcarenite towards the upper part of the succession. It is also characterised by the presence of significant

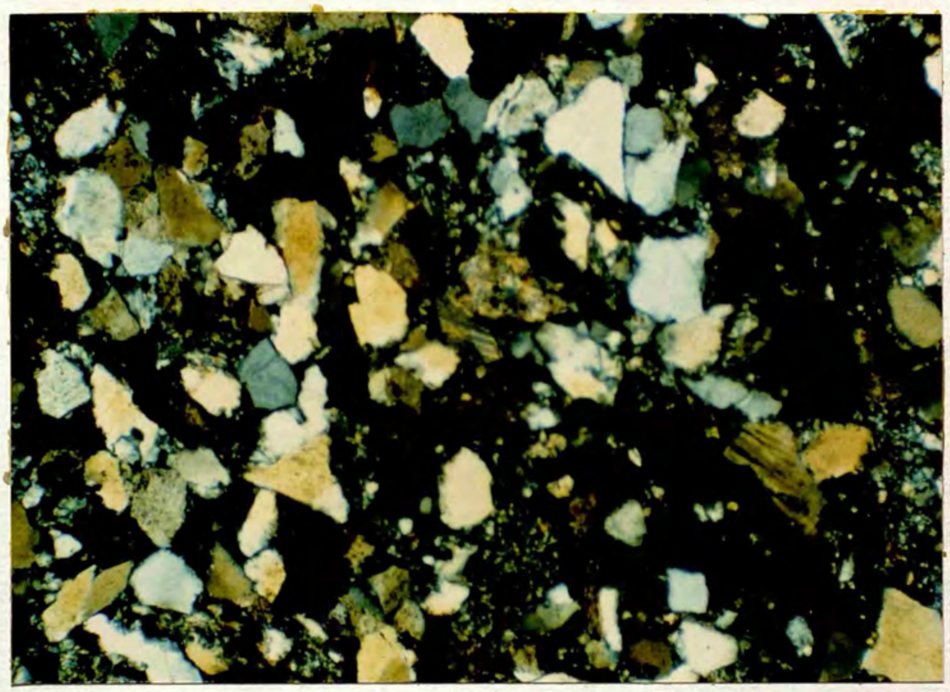


Plate 4.1 Photomicrograph of quartz-rich lithic arenite of Kolo Beds (83 TO 42.2), showing significant amounts of lithic fragments subsidiary to terrigenous quartz. The rock fragments include metamorphics, peridotites, altered basalt, chert and limestones. Note the poor sorting and immaturity of the arenite. Crossed polars, 40X.

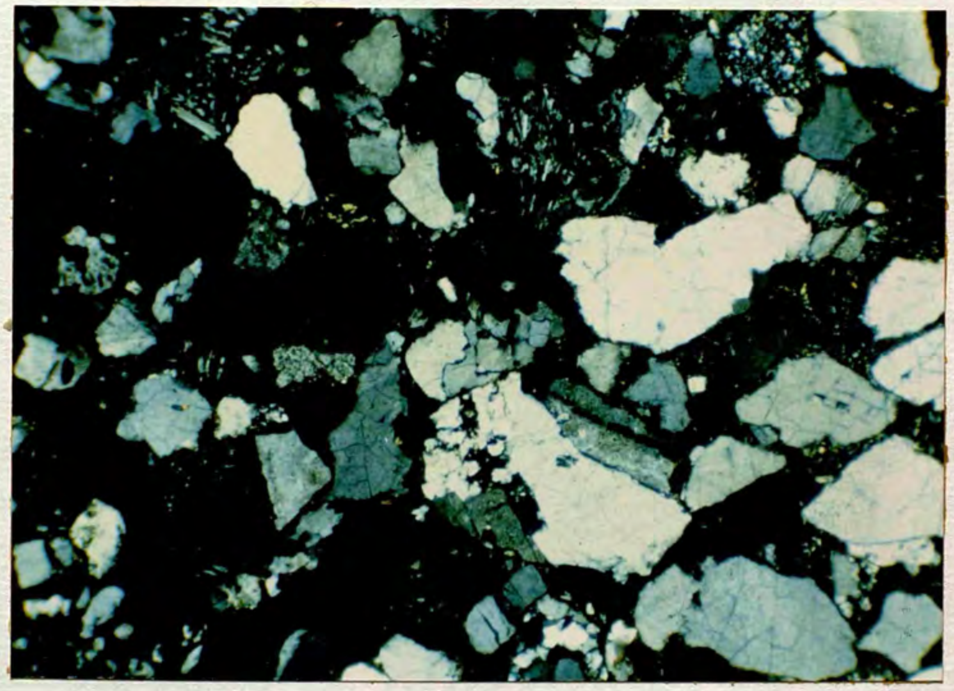


Plate 4.2 Photomicrograph of quartz-rich lithic arenite of Kolo Beds (83 TO 195.4), showing the occurrence of subrounded quartz detritus and presence of altered volcanic fragments and rounded metamorphic fragments (bottom centre) and siliceous mudstone (near top right corner). Crossed polars, 40X.

amounts of organic matter giving rise to the dark and brownish colour of the rocks.

The arenites are well-bedded, with bed thickness ranging from 4 to 70 cm. Parallel laminae occur in most arenite beds; mudstone and/or carbonaceous silty shale chips of angular and elongated shape up to 10 cm long occur in the lower portion of the thicker arenite and/or pebbly arenite beds. Current ripple lamination or current foresets occur in some bed of arenites. Most beds show very sharp basal and upper contacts.

Pebbly arenites usually occur in the lower part of the facies, and the grains gradually decrease in size to the overlying coarse arenites. In outcrop, they are dark, yellowish to brownish in colour, compact and lithified; thickly bedded, with beds usually more than 50 cm thick.

Most of the thicker beds of arenite are structureless, but some may show a weak grading. Overall, the sedimentary features present suggest that this succession was deposited by turbidity currents.

Generally the Kolo Beds dip towards the south, which contrasts with the northnorthwestward steep dip and imbrication of the older units such as the Salodik Limestones, and the Kapali Beds and the highly deformed chert sequence.

The calcarenites and calcareous silty shales contain abundant planktonic foraminifera and minor benthic foraminifera. Benthic macroinvertebrates also occur in some beds of calcarenite. The matrix of pebbly mudstones contains predominantly planktonic foraminifera, but the fragments may be rich in benthic foraminifera and/or benthic macroinvertebrates and planktonic foraminifera as well. This features indicate that they were derived from inner shelf.

In thin section, the arenite facies is typically quartz-rich lithic arenite consisting largely of quartz detritus, and subsidiary lithic fragments and detrital grains of feldspar, and calcite detritus in the calcarenite. In the upper portion of the succession, carbonates and microfossils become more abundant in the calcarenite beds. Additional terrigeneous grains present in small amounts include muscovite, hornblende, chlorite and opaque mineral euhedra.

These grains are set in a matrix consisting of mixed clay mud, cryptocrystalline quartz with cement of silica and iron oxides. The calcarenite has a matrix consisting largely of micritised calcite and subsidiary clay mud. Locally, sparry calcite occurs, probably as a secondary filling of the pore spaces.

The QFL diagram (Fig. 4.5.2) clearly shows that this facies is a recycled orogenic sand in the sense of Dickinson (1979), indicating that fragments were derived from subduction complexes or collision orogens.

Quartz detritus is polycrystalline, angular to subrounded in shape, with size grades up to 0.4 mm across. The larger grains are monocrystalline, some of them show a wavy or undulatory extinction and strain features, which point to pre-depositional plastic deformation. Some of the grains show marginal embayment due to corrosion. The cryptocrystalline quartz occurs as small grains and in the matrix. Some of them are probably derived from recrystallised siliceous mudstone and/or chert (e.g. 83TO195.4). The quartz detritus makes up to 60% of the rocks (e.g. 83 TO 42).

Detrital grains of feldspar are subhedral and subrounded to prismatic in shape, with size grades up to 0.8 mm long. K-feldspar and plagioclase are both present. Varieties of K-feldspar include microcline and orthoclase. The microcline grains are characterised by cross-hatched twinning. Plagioclase shows typical albite and carlsbad twinning and consists largely of labradorite (An66-76) and subsidiary andesine (An36-48) and minor albite-oligoclase (An4-11). Some of the feldspar grains show marginal and/or zonal alteration and are replaced by sericite, clays and carbonate. Feldspar detritus may form up to 7 % of the rocks (e.g. 83 TO 158.2).

Most of the arenites contain detrital muscovite, which may constitute up to 1% of the rocks (e.g. 83TO165.1). The larger grains are prismatic in shape with size grading up to 0.3 mm long and show a wavy or undulatory extinction due to plastic deformation. Sericite, chlorite and opaque minerals are present in very minor amounts (less than 1%).

The rock fragments are typically polygenetic grains, consisting of intermediate to acidic volcanics, basic and ultrabasic rocks, metamorphics, and sedimentary rocks, including siliceous mudstone, calcilutite, chert and carbonaceous silty shales. Lumps of coal and carbonaceous silty shales fragments up to 10 cm long occur in some beds of arenite. Lithic fragments may make up to 25% of the rocks (e.g. 83 TO 165).

The intermediate volcanic fragments are characterised by a microlitic texture. Rhyodacitic fragments contain phenocrysts of altered feldspar set in a groundmass of colourless glass carrying microlites of feldspar and pyroxene and crowded with acicular crystallites (e.g. 83TO195.4B). The volcanic fragments are subrounded in shape with size grading up to pebble size, and most of them are highly altered (e.g. 83 TO 43B).

The basic-ultrabasic fragments include basalt, dolerite, peridotite and serpentinite. Most of the fragments are partially or completely altered. The dolerite fragments show ophitic texture and the serpentinite fragments show mesh structure due to derivation from olivine. Basalt contains altered labradorite phenocrysts and a matrix of glass with microlites of highly altered plagioclase and pyroxene (e.g. 83 TO 156).

The metamorphic fragments include mica-schist, phyllite and meta-quartzite. A few grains showing gneissic texture are present in some rocks (e.g. 83 TO 195.4B). Metamorphic grains may be present up to 10% of the rocks (e.g. 83 TO 171.2). They are subrounded in shape with size grades up to 0.5 mm across. The siliceous mudstone fragments are subrounded and elongated in shape with size grading up to 1.5 mm long. Some of the grains show ghosts of microfossils.

Chert fragments are typically angular in shape with size grades up to 0.5 mm long and may constitute up to 5% of the rocks (e.g. 83 TO 171.2). Some grains contain numerous ghosts of radiolaria. Criss-cross quartz veinlets occur in some grains. Some of the grains were formed from recrystallised siliceous mudstone or argillite. Calcilutite fragments consist of micritised lime mudstone and wackestone and are present in small amounts (5%). Some of them contain scattered calcispheres of microfossils, including radiolaria. They are subangular and elongated in shape with size may grade up to 1.5 mm long.

Carbonaceous fragments of oval and elongated shape with size grades up to 10 cm long occur in some beds of arenite. They are commonly found within the lower part of the beds of arenite, and may show fabric imbrication of their long axes. The fragments are derived from carbonaceous silty shale and redeposited by turbulent currents in the Ta or Tb intervals. They consist primarily of plant remains and are typically very dark in colour (e.g. 83 TO 172.1). Most of the carbonaceous fragments are pyritised.

Fossil fragments consist largely of calcispheres of planktonic foraminifera and minor micritised and abraded skeletal debris, and may constitute up to 20% of the rocks (e.g. 83 TO 158). The abraded nature of the fragments points to reworking and redeposition of the skeletal debris.

These detrital grains and rock fragments are set in a

Fig. 4.4 Detailed section of parts of the Kolo Beds in the Gumbangi River (83T)165), showing very high ratio of sand to shale (8 : 1) and very high proximality index (PI = 80%).

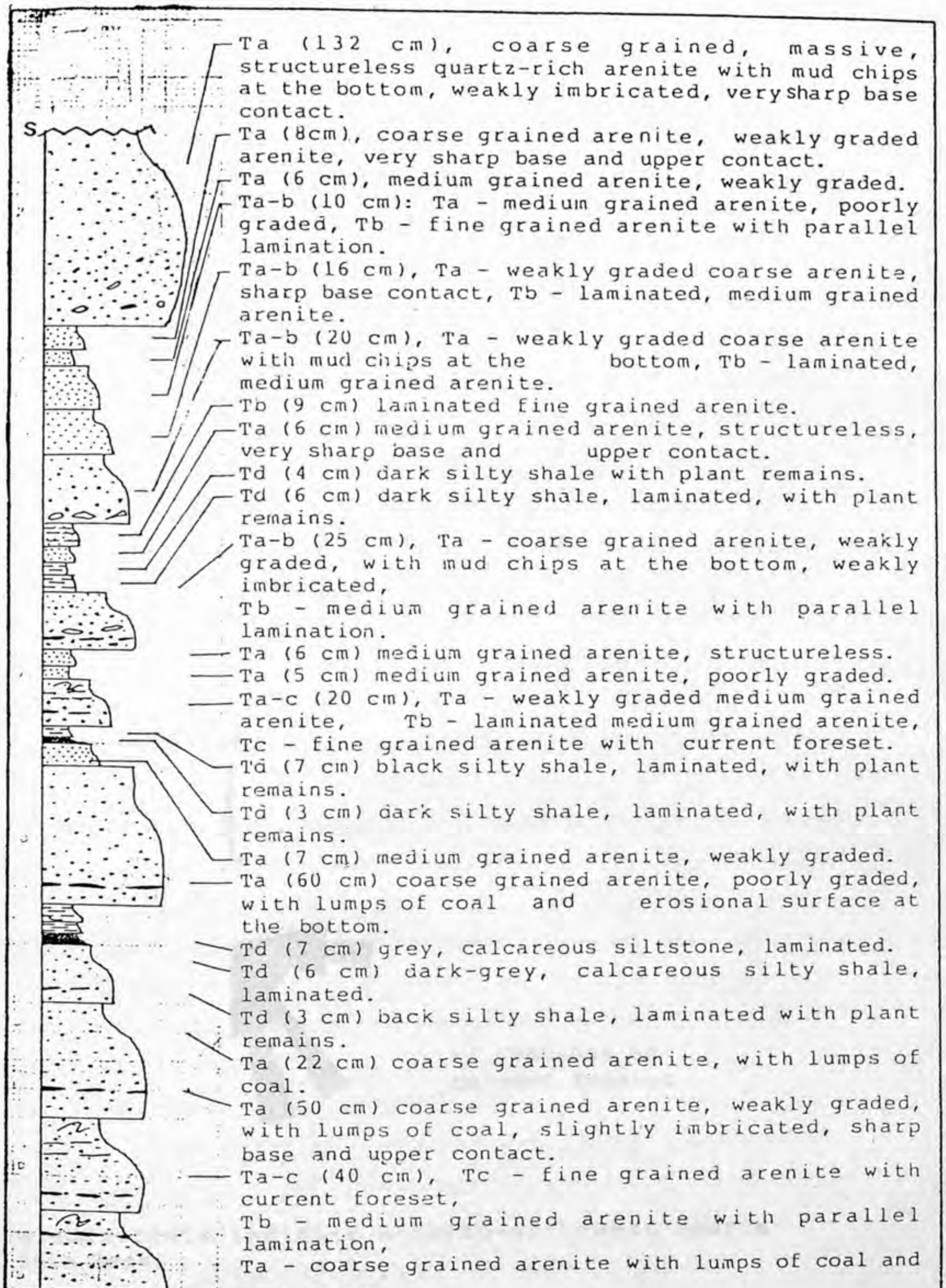


Fig. 4.5.1 Paleocurrents indicate a north-northwest source area for the Kolo Beds.

matrix of mixed cryptocrystalline quartz and clay mud with a cement of silica and iron oxides. Carbonate cement is also locally present.

The arenites and calcarenites are compositionally and texturally immature which points to relatively rapid erosion, transportation and deposition by turbulent currents. The rock fragments and detrital grains compositionally strongly indicate that the arenites are derived from source area consisting of two different tectonic terrains, i.e. continental margin sequence juxtaposed with ophiolite belt. This feature may suggest, further, at least some parts of the collision complex was uplifted during deposition of the Kolo Beds.

(ii) **Argillaceous Lithofacies**

The argillaceous rocks are well bedded with thickness ranging from 4 to 10 cm, but a few beds of 15 cm thick are present. Most beds show thin parallel lamination with thickness of less than 2 cm. Current ripple lamination or current foreset may be present in some beds. A few beds show flame structures. Erosional surfaces occur at the top of some beds. Argillaceous rocks occur as alternating beds the arenites and calcarenites and in the upper part of the turbidite beds, usually occurring in Tc and/or Td intervals. The basal contact is usually sharply defined in the alternating beds, but is gradational into the underlying arenite in the turbidite beds.

In thin sections, the facies consists largely of argillaceous marlstone and minor carbonaceous silty shale. The marlstones are yellowish wackestone and lime mudstone. The wackestone contains large numbers of micritised calcispheres of planktonic foraminifera, abraded skeletal debris and minor quantities (less than 1%) of quartz detritus. Fossil grains may constitute up to 40% of the rocks (e.g. 83 TO 50A). The matrix consists largely of lime mud and minor cryptocrystalline quartz and clay. The lime mudstones are dominated by lime mud with a few scattered calcispheres of foraminifera (e.g. 83 TO 46B).

The carbonaceous silty shales consist of detrital grains of quartz, feldspar, muscovite and plant remains; calcite detritus may be present in some rocks. Most of the grains are angular and grade up to silt size. The plant remains are pyritised, very dark in colour. The laminae in the carbonaceous silty shales are defined by the alternation of very dark or black laminae containing large amounts of plant remains, and the light-grey quartz-rich laminae (e.g. 83 TO 172.2). The occurrence of carbonaceous silty shale laminae suggests that the plant remains were reworked and redeposited by low velocity currents.

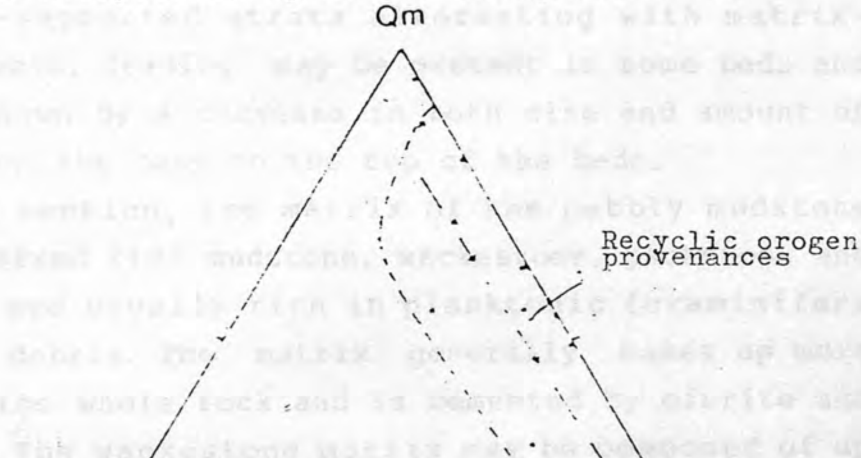


Fig. 4.5.2 Triangular QmFL plot (Dickinson, 1979),showing that the Kolo Beds are derived from recycled orogen provenances. Dashed-line with arrows indicates increasing ratio of oceanic/continental fragments.

(iii) Pebbly Mudstone Lithofacies

Pebbly mudstone is typically a matrix supported coarse clastic sediment. In outcrops, the rock is thickly bedded with bed thickness up to 10 m or more. The base contact of each bed is sharply defined and usually is marked by an erosional surface cut into the underlying layers. The rock consists of poorly sorted fragments of pebble to block size set in a matrix of mixed lime mud, silt and/or calcarenite. The fragments range in size from sand to 10 cm long, angular or elongated in shape, and are randomly oriented. Some beds may show stratification of tens of centimetres thick, as indicated by the occurrence of fragment-supported strata alternating with matrix-supported strata. Grading may be present in some beds and usually is shown by a decrease in both size and amount of fragments from the base to the top of the beds.

In thin section, the matrix of the pebbly mudstone consists of mixed lime mudstone, wackestone, packstone and calcarenite and usually rich in planktonic foraminifera and skeletal debris. The matrix generally makes up more than 70% of the whole rock and is cemented by micrite and iron oxides. The wackestone matrix may be composed of up to 40% calcispheres of microfossils, largely of planktonic foraminifera (e.g. 83 TO 50A) and quartz detritus of silt size which forms up to 8% of the rock; a few feldspar grains of silt size are also present (e.g. 83 TO 50C).

The packstone matrix contains fossil fragments, including calcispheres of planktonic foraminifera and skeletal debris up to 80% (e.g. 83 TO 44) and minor quartz detritus and chert; a few glauconite pellets are also present, probably were derived from the Mesozoic to Early Tertiary continental margin sequence described previously in Chapter 2. The lime mudstone matrix contains very minor and scattered calcispheres of microfossils, usually less than 2% (e.g. 83 TO 46).

The lithic clasts consist of a variety of fragments including limestones, sandstones, siliceous and lime mudstones, chert, and serpentinised peridotites and generally range from granule to pebble size, but some clasts may be up to tens of centimetres long; a few fragments or blocks of more than 1 metre long may be present. The fragments are typically angular and elongated in shape and are extremely unsorted.

The limestone fragments consist of packstone, wackestone and grainstone, which contain both benthic and planktonic foraminifera, similar to those of the Salodik Formation (e.g. 83 TO 168.3; 83 TO 44). The sandstone fragments consist of calcarenite,lithic arenite and quartzose arenite. The quartzose arenite and some of the lithic arenite are texturally and compositionally similar to those of the Kapali Beds, and the calcarenites are similar to those of the Kolo Beds described previously (e.g. 83 TO 183.3). The siliceous mudstone fragments (e.g. 83 T0 159) contain very minor (less than 10%) detrital grains of silt size, including quartz, feldspar, biotite, amphibole and sericite. A few scattered recrystallised ghosts of microfossils are also present in the mudstone. The calcilutite fragments contain up to 5% micritised and infilled sparry calcite calcispheres of microfossils (e.g. 83 TO 46). The chert fragments contain numerous ghosts of radiolaria (e.g. 83 TO 162.1).

The serpentinised peridotite fragments range in size from granule up to 10 cm, angular in shape. Some of the fragments are foliated. In thin section the serpentinised peridotites consist largely of fibrous antigorite (e.g. 83 TO 195.4). Basalt fragments are also present in some beds.

C. Biostratigraphy

The Kolo Beds contain abundant planktonic foraminifera, while benthic foraminifera are present in very small amounts. Samples collected from 15 localities were examined by Purnamaninsih-Siregar, Sudiono, Budiman and Dr.Darwin Kadar, Paleontological Laboratory of the Geological Research and Development Centre (GRDC), Bandung, Indonesia. Age determination is based on the foraminiferal biostratigraphic zonations by Blow (1964), Postuma (1974) and Hainforth et al.,(1975).

In the stream course of the Kapali River, the matrix of the pebbly mudstone, calcarenite and calcareous silty shales (e.g. 83 TO 145, 83 TO 146, 83 TO 155, 83 TO 168, 83 TO 168.2, 83 TO 168.3, 83 TO 169.1) contains planktonic foraminifera, including Orbulina universa, Globigerinoides trilobus (Reuss), Globigerinoides extremus Bolli & Bermudez, Globorotalia menardii D'Orbigny, Globorotalia culturata, Globiquadrina sp., Amphistegina sp. and Sphaeroidinellopsis subdehiscens (Blow) of Late Miocene to Pliocene (N16-21) age.

In the stream course of the Gumbangi River, the calcareous silty shales, calcarenite and in the wackestone-dominated matrix of the pebbly mudstone (e.g. 83 TO 164.2, 83 TO 165, 83 TO 165.1, 83 TO 166.2, 83 TO 167) contain abundant planktonic foraminifera, including Orbulina suturalis Bronniman, Globigerinoides trilobus (Reuss), Globigerinoides sp., Globorotalia menardii D'Orbigny, Globorotalia acostaensis Blow, Globiquadrina altispira (Cushman & Jarvis), Sphaeroidinellopsis subdehinscens Blow and Operculina sp. of Late Miocene to Pliocene (N16-21) age.

In the stream course of the Pangkape River (e.g. 83 TO 157.3, 83 TO 258.3, 83 TO 1159.2, 83 TO 182), the marly sandstones, calcareous silty shales and matrix of the pebbly mudstone contain planktonic foraminifera,including Orbulina universa D'Orbigny, Globorotalia menardii D'Orbigny, Globorotalia situlata, Globigerinoides trilobus (Reuss), Globorotalia immaturus Bronniman, Globorotalia sacculifer (Brady), Globorotalia obliquus Bolli, Globoquadrina altispira (Cushman and Jarvis), Globigerina venezuelana Hedberg of Late Miocene to Pliocene age (N15-19).

The age of the Kolo Beds, based primarily on the planktonic foramninifera, is Late Miocene to Pliocene (N16-21).

Benthic macroinvertebrates occurring in some beds of calcarenite were examined by Dr. H.G. Owen, Mr C.P. Nuttall and others, of British Museum, Natural History. Mr. Nuttall (person. com., 1985) identified bivalves and gastropods occurring in the pebbly mudstones, which include Solecurtus sp., Antiogona sp., Pinna sp., and Pitar sp. He concluded that the exact age of the faunas cannot be determined but the age is certainly Tertiary.

The occurrence of rare benthic foraminifera, including Lepidocyclina sp. in the matrix of pebbly mudstones suggests that the age is older than that given by most of the planktonic foraminifera described above. The possibility of reworking of the benthic foraminifera is also suggested by their scarcity in the sediments.

D. Stratigraphic Relationship Between Lithofacies

The geological traverses made along the stream courses of Kapali, Gumbangi, Bahooti and Pangkape Rivers show that the Kolo Beds gradually decrease in both grain size and bed thickness from NNW to SSE, which signifies the gradual facies changes both laterally and vertically. This feature is also well-pronounced by a gradual decrease in both the sand-shale ratio and the proximality index (Pl) from NNW to SSE. These features all suggest that the direction of transport was trending from NNW to SSE.

In the northern part, the sequence is dominated by arenite intercalated with pebbly mudstones and minor silty shales. The pebbly mudstones were probably deposited by debris flows, originally as lenses or channel fills occurring in the lower part of the succession. The coarse clastic sequence is gradually overlain by arenite and calcarenite, alternating with silty shales, which formed the upper portion of the unit and occupies the southern part of the area. The upper sequence was predominantly deposited by turbidity currents and partly, particularly the thicker and structureless arenite beds, by fluidised and/or grain flows.

The facies pattern of the Kolo Beds resembles submarine fan deposits (Walker, 1981), although the distal portion of the unit does not occur in the Kolo Atas area.

E. Discussion and Interpretation

The Kolo beds display the sedimentological features typical of recycled orogenic clastic rocks. Sedimentary features of the sediments strongly suggest that they are typically products of sediment gravity flow deposits, i.e. the arenite and calcarenite lithofacies were deposited by turbidity currents, while the pebbly mudstone were deposited by the mass-movement of subaqueous debris flows. The thicker and structureless arenite beds are more likely to be deposits of fluidised and/or grain supported flows.

The nature of the pebbly mudstone suggests that some of the rocks might have originated as olistostromes, particularly those rocks which contain fragments derived from the older rocks (i.e. from Salodik Formation, Kapali Beds and the ophiolite suite). As Abbate et al.,(1970) pointed out, the term 'olistostrome' was originally given to a wide spectrum of redeposited clastics, but has since been redefined to apply to chaotic deposits emplaced by debris flows and related mass gravity processes which are composed of extra-formational material or which contain exotic clasts derived from the older rocks. The pebbly mudstone of the Kolo Beds, however, also contains significant amounts of clasts of intra-formational origin, including calcarenite and lithic arenite.

The origin of the pebbly mudstones is considered to be due primarily to mass-gravity transport, under the direct influence of the force of gravity and where the fragments are mixed with water, it is the sediment that moves the fluids (Rupke, 1982). The process generally takes place spasmodically, that is, intermittently and catastrophically by the rapid downslope displacement of large masses of sediment. Moore (1961) and Morgenston (1967) explained that mass-gravity transport is closely related to the mechanics of sediment failure and geological conditions that trigger failure. When sediment is deposited on a slope surrounding a basin, it will move downslope only when the shear stress exerted by the force of gravity exceeds the shear strength of the sediment. The shear strength is a function of the cohesion between grains plus the intragranular friction. A very small angle of slope (ranging from 0.5 to 6 degrees) is sufficient for mass sediment transport (Roberts, 1972).

The materials forming the sediments, including rock fragments and detritus, strongly suggest two different suites of provenance. The occurrence of intermediate-acidic volcanic fragments and terrigenous grains of quartz, K-feldspar, muscovite and biotite and some of the metamorphic fragments, at least the gneissic grains, strongly suggest that the source rocks formed part of a continental basement complex, or more likely, these materials were derived from recycling of continental shelf-derived sediments. This corresponds to the Mesozoic and Early Tertiary continental margin sediments including the Triassic Lemo Beds and Jurassic Kapali Beds described previously in Chapter 2.

On the other hand, the presence of the basic and ultrabasic fragments, plus detrital grains of basic plagioclase, including labradorite and andesine, strongly points to an ophiolite suite provenance. The presence of radiolarian chert fragments suggest pelagic cover to the ophiolite suite provenance. The radiolarian calcilutite, however, could have been derived from either continental shelf deposits or from the pelagic part of the ophiolite complex.

The two distinctive types of provenance of the Kolo Beds strongly suggests that the succession was derived from the collision complex between the Banggai-Sula Platform and the Eastern Sulawesi Ophiolite Belt or subduction complex. The Late Miocene to Pliocene Kolo Beds, therefore, can be classified as Neogene post-orogenic sediments. An important conclusion is that the collision of the Banggai-Sula Platform against the Eastern Sulawesi Ophiolite Belt occurred in pre-Late Miocene time. The age of collision will be further discussed in Chapter 5.

The most significant sedimentological feature of the Kolo Beds is a very high ratio of sand to shale, indicating a proximal depositional setting. The distal portion of this succession is not seen in the type section, in Kolo Atas area. However, the exposures suggest that sand/shale ratio decreases toward the southwest and south, indicating that the distal portion of the succession might occupy the sea floor of the Tolo Bay-Peleng Strait.

The predominance of planktonic foraminifera indicates that the Kolo Beds were deposited in an open basin and in relatively deep water. This might be regarded as a foreland trough separating the Banggai-Sula Platform to the south from the Eastern Sulawesi orogenic complex developed in Middle Miocene time, or possibly as trench-fill or slope basins on the hangingwall of the accretionary complex.

4.3.2. BIAK CONGLOMERATES

A. Definition

A succession of coarse clastic rocks dominated by conglomerates, exposed along the Biak-Poh road, just to the north of Biak village is informally named the Biak Conglomerates. The unit overlies the Salodik Limestones unconformably, and is unconformably overlain by Quaternary coralline reef deposits.

B. Synonymy and derivation:

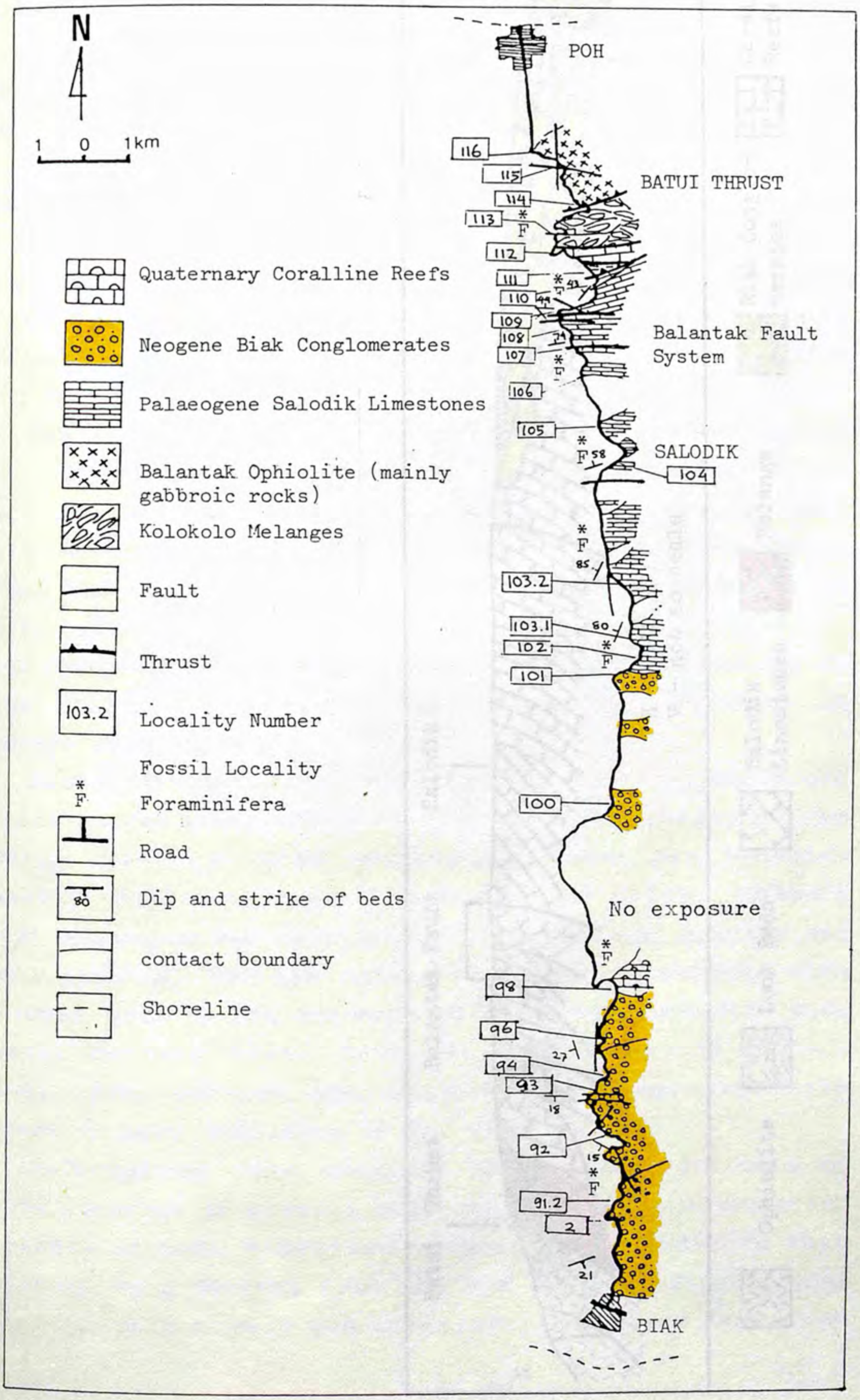
The unit was named the Biak Formation by Rusmana et al.(1984).The name is derived from Biak village, where the best exposures occur (Fig.4.6). The Biak Conglomerates, were included in the upper part of the Celebes (Sulawesi) Molasse of Sarasin and Sarasin (1901) and form part of the Bongka Group (Simandjuntak et al., 1983; Rusmana et al., 1984).

C. Description:

The Biak Conglomerates are typically coarse clastic sediments and consist largely of grey and dark grey conglomerates, and subsidiary arenites with minor silty marl. The rocks are well-bedded with bed thickness ranging from 3-5 cm in the silty marls and silty shales up to 4 m in the conglomerate beds. The arenite beds usually range in thickness from 5 to 70 cm, and the pebbly arenites may attain a thickness of 30 cm to nearly 1 metre. The basal contact of the bedding is sharply defined and the upper contact may either be sharp or gradational.

The lower part of Biak Conglomerates shows a fining and thinning upwards sequence. The pattern usually starts with conglomerates at the base, which grade up into pebbly

Fig. 4.6.1 Geological traverse map of the Biak-Poh section, showing the occurrence of Luok Beds and Salodik Limestones juxtaposed with the Balantak Ophiolite and the overlying coarse clastic rocks (i.e. Biak Conglomerates).



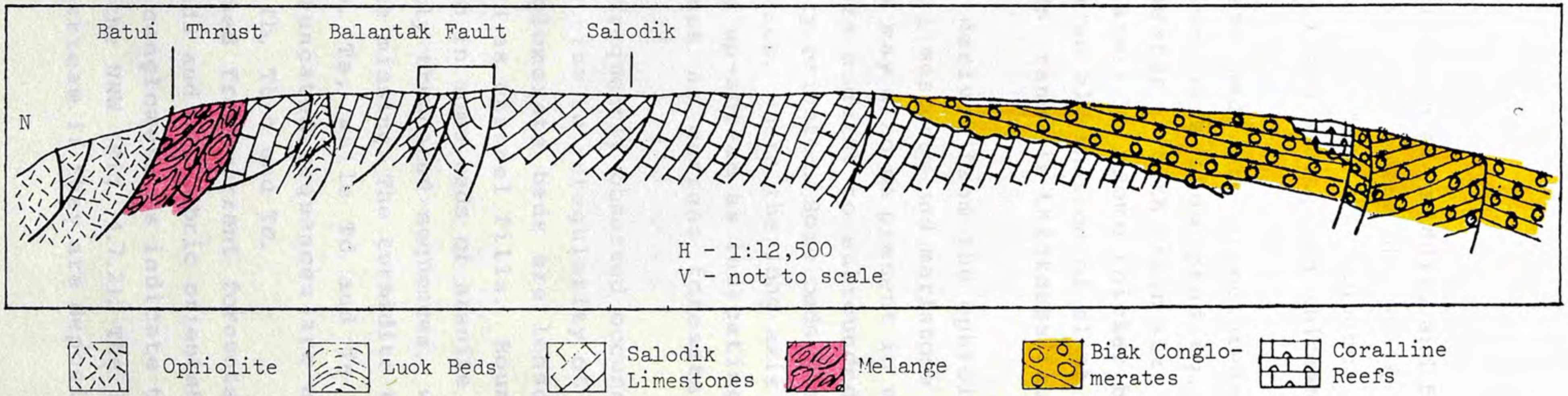


Fig. 4.6.2 Line section through central part of Biak-Poh Section showing the ophiolite thrust over an imbricated carbonate platform sequence of Salodik Limestones and the post orogenic clastic Biak Conglomerates.

arenites or very coarse arenites in the middle, and fine grained arenites or silty marls or silty shales in the upper part. While in the upper part of the sequence the Biak Conglomerates change to a coarsening and thickening upwards.

Most of the conglomerate beds are structureless, massive and lithified. Some beds show grading, as indicated by the decrease upwards in both grain size and the proportion of clasts (Plate.4.3D). Some thicker beds may show stratification with an alternation of clast and matrix supported strata. Beds range in thickness from 2 to 15 cm.

The clasts are largely derived from the ophiolite suite, but include subsidiary limestones and marlstone and minor chert; quartz detritus may also be present in very small amounts. The clasts are angular to subrounded in shape, and are mostly randomly oriented. Some beds show a weakly oriented fabric of clasts, where the long axis (b-axis) imbricated and dipping up-stream as indicating by other current indicators,such as current foresets in adjacent arenite beds.

Erosional surfaces are frequently observed occurring in this succession, and give rise to irregularity of the bedding surfaces. Some conglomerate beds are lensoid, probably they were deposited as channel fills. Bouma's (1962) sequences are developed in some beds of arenite and pebbly arenite, but are mostly truncated sequences, with the upper part of the sequence missing. The turbidite beds usually contain Ta-c, Ta-b, Ta, while Td and Te are absent. Base-cut-out and truncated sequences are also present in some beds,such as Tb, Tb-c and Tc.

Paleocurrent data obtained from current foresets of the Tc interval in arenite beds and the fabric orientation of clasts of some organised conglomerates indicate that the rocks were derived from the NNW (Fig. 4.7.2). The long axes of clasts always dip upstream if they are deposited

Plate 4.2.1 Photograph of outcrop of the Biak Conglomerates in 83 TO 93.1, showing organised conglomerate with size generally less than 5 cm in diameter.

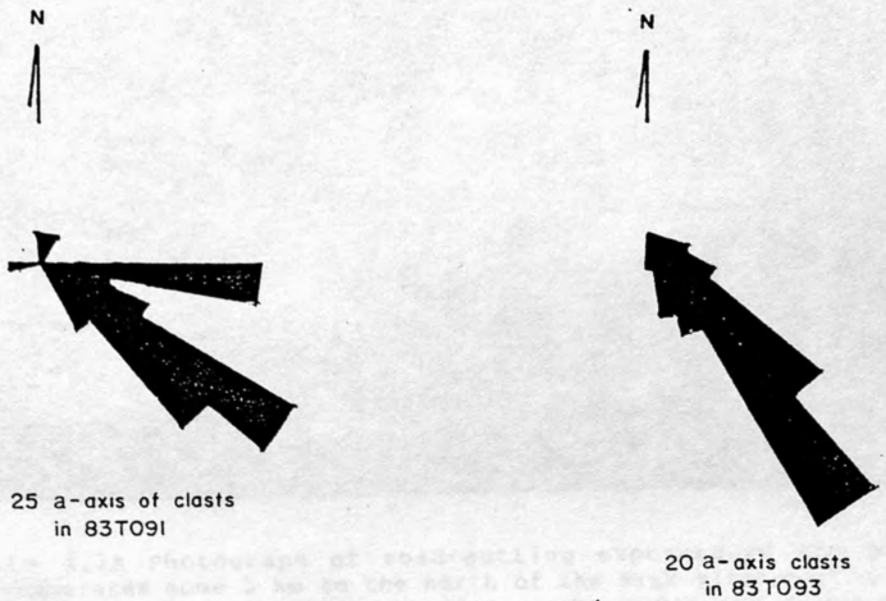


Fig. 4.7.2 Paleocurrents indicate a north-northwest source area of Biak Conglomerate.

Plate 4.3A Photograph of road-cutting exposure of the Biak Conglomerates some 5 km to the north of the Biak village (i.e. 83 TO 96), showing conglomerate with subrounded and elongated clasts set in a matrix of calcareous sandy mud. Note the weakly imbricated nature of the clasts in the lower portion of the bed, and the slightly graded clasts to pebbly conglomerate at the top of the bed.

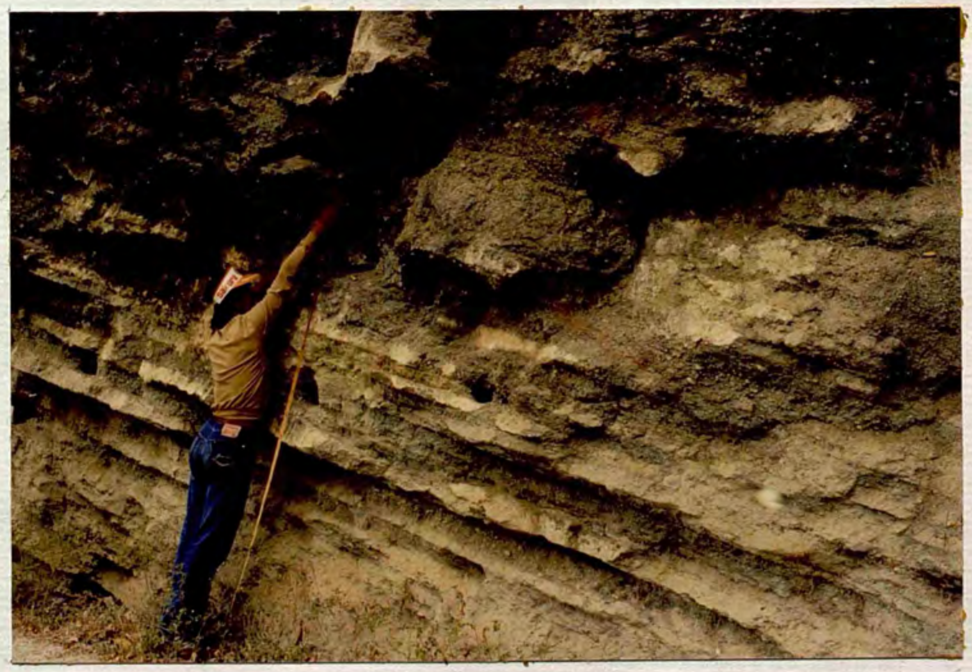


Plate 4.3B Photograph of road-cutting exposure of Biak Conglomerates, some 4 km to the north of the Biak village (i.e. 83 TO 93), showing a sequence of well-bedded calcareous lithic arenite with intercalations of thinly bedded marlstone.

by turbulent currents, in contrast to clasts which are rolled in a beach or river, resulting in the long axes lying perpendicular to the current. The NNW-derivation of these rocks fits the present physiographic configuration of the region, which is not considered to have greatly changed since late Neogene time.

The term 'organised' conglomerate is used in the sense of Walker (1978) for conglomerates which show a regular orientation of clast fabric, such as imbrication and stratification. 'Disorganised' conglomerate has a typical randomly oriented clast fabric. Organised and disorganised conglomerate essentially contain over 50% clasts. Paraconglomerate is typically a matrix-supported conglomerate and contains less than 50% clasts and it can be either organised or disorganised.

The rocks are much less-deformed than the underlying rock units, but are faulted in many places and are gently tilted, dipping towards SSW. Joints and fractures only occur locally, particularly in fault zones.

The unit crops out in the Biak river valley and is well-exposed for 4 km along the Biak-Poh road, just to the north of Biak village (Fig. 4.3). The exposure forms a north-south trending U or V shape body, which now coincides with the valley of Biak river. The maximum thickness of the unit may not more than 500 metres.

For purposes of description and interpretation the Biak Conglomerates are divided into:

(i) coarse-clastic and
(ii) fine-clastic lithofacies.

(i) <u>Coarse-clastic</u> <u>lithofacies</u>

This facies consists largely of disorganised conglomerate and subsidiary organised, paraconglomerates and pebbly arenites.

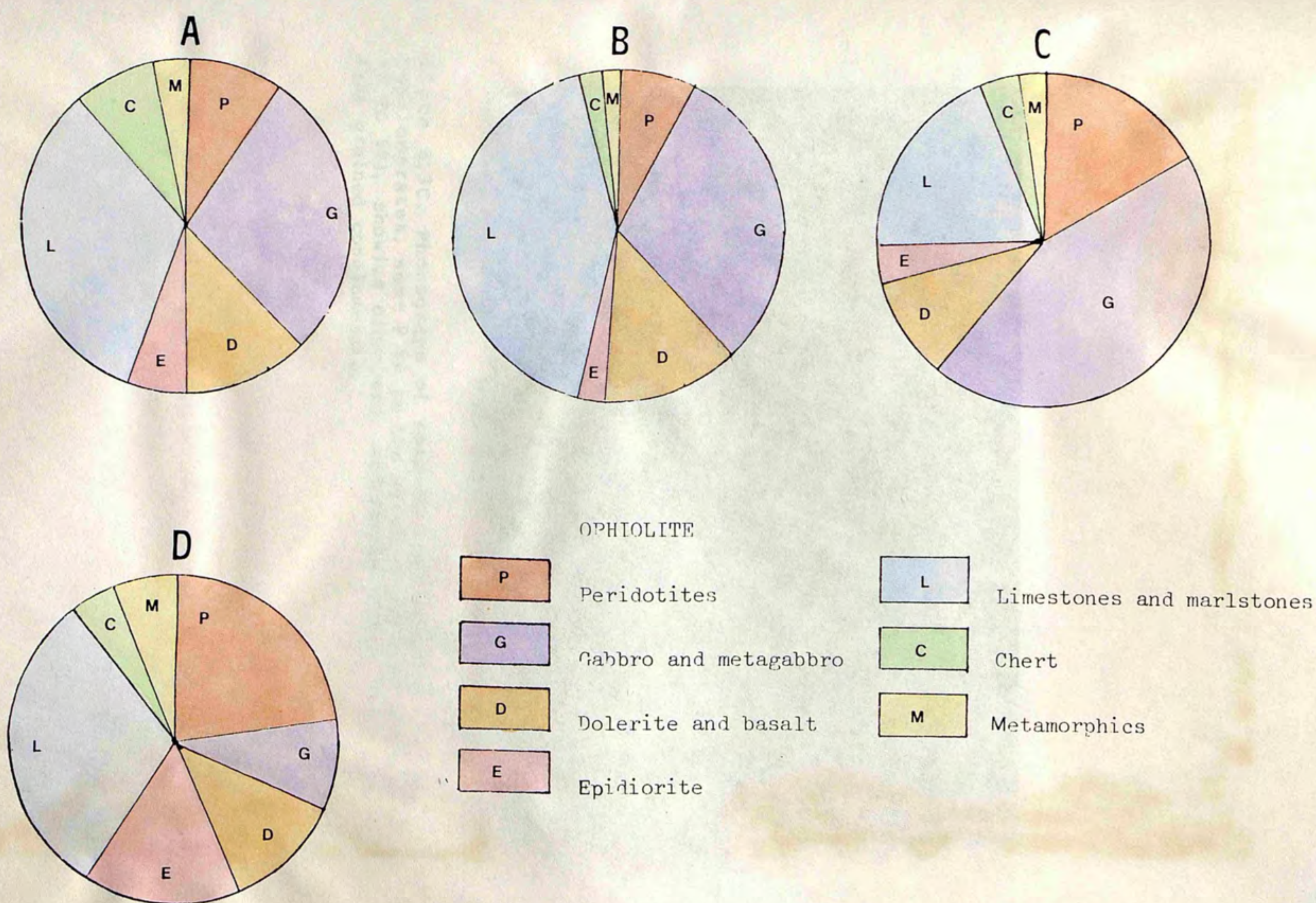


Fig.4.7.3 Diagram showing composition of clasts of conglomerates in the Biak Conglomerates. (A:83T02, B:83T091.2, C:83T092, D:83T093).

Plate 4.3C. Photograph of road-cutting exposure of the Biak Conglomerates, some 8 km to the north of the Biak village (i.e. 83 TO 99), showing organised, moderately sorted, and relatively fine grained conglomerate.

On the basis of counting the clasts present in the conglomerates within 1 m.sq., the clasts of the conglomerates consist largely of ophiolite derived fragments (up to 80%), limestone and marlstone (up to 30 %), and chert (up to 3%) in both the organised and disorganised conglomerates (Fig. 4.8).

The clasts are angular and subrounded in shape; tabular clasts are also present in some beds. They range in size from granule up to 25 cm across; but generally in the range of 5 to 15 cm. Chert, epidiorite and metamorphic fragments are commonly of smaller size (less than 5 cm long) and angular in shape. The clasts are commonly moderately sorted, and have a poor to moderate sphericity. Some of the organised conglomerates show a weakly oriented fabric of clasts, where the long axis (b-axis) is parallel and dipping up stream as indicated by current foresets occurring in the adjacent arenite beds. These clasts are set in a matrix of mixed sand, mud, clay and locally carbonates.

As described previously, the clasts in the conglomerates consist largely of ophiolite fragments, included gabbro, dolerite, basalt and minor diorite, peridotites, serpentinite, epidiorite and metagabbro. In thin sections the ophiolite fragments are highly altered. In the mafic rocks, the plagioclase cystals in both matrix and phenocrysts are partially or totally replaced by sericite, carbonate, zeolite and albite. Albitisation of plagioclase commonly occurs in gabbro and dolerite. Gabbro fragments contain plagioclase of labradorite composition (An 60-64) up to 40% (e.g. 83TO91.4), which are subhedral, and tabular in shape, up to 3 mm long. Plagioclase shows typical carlsbad and albite twinning.

The dolerite fragments often show ophitic plagioclase-pyroxene intergrowth texture and contain plagioclase of labradorite (An 55-70) and locally andesine composition (An45), up to 50% of the rock (e.g. 83TO91.1),

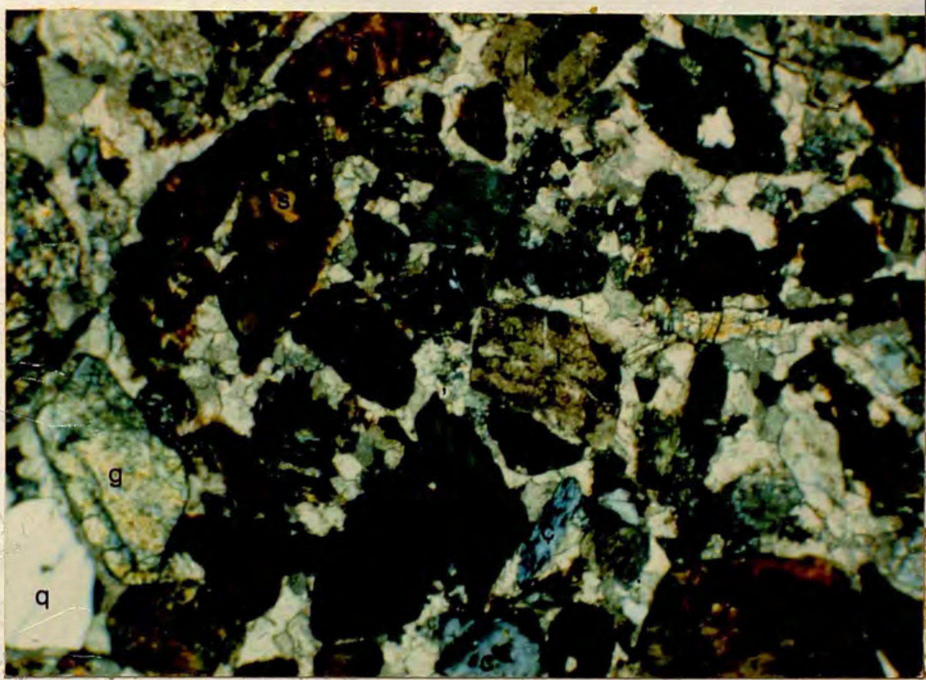


Plate 4.4A Photomicrograph of lithic arenite of Biak Conglomerates (83 TO 95), showing angular to subrounded lithic fragments including serpentinised peridotite (s), gabbro (g), schists (m), chert (c), limestone (l) and detrital grains of quartz (q), pyroxene (p) and minor feldspar. Secondary sparry calcite is also present as cement. Crossed polars, 40X.

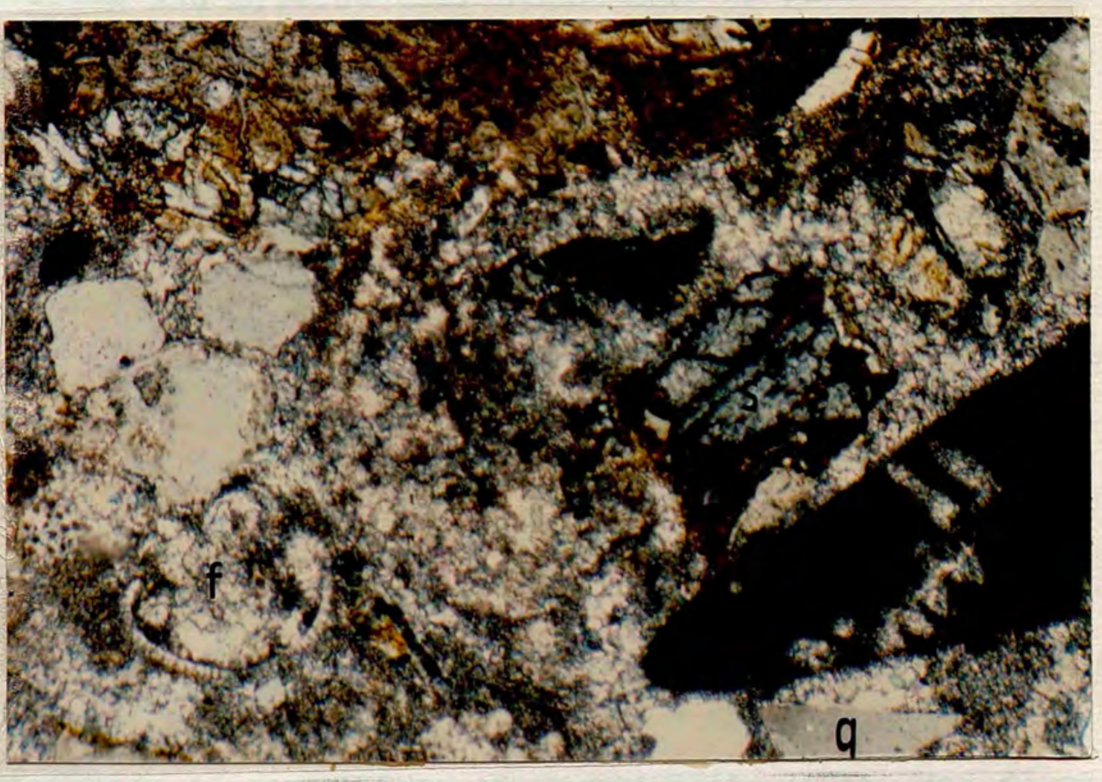


Plate 4.4B Photomicrograph of calcareous lithic arenite in the Biak Conglomerates (83 TO 3), showing planktonic foraminifera (f) in the left bottom corner, and coralline algae (a) in the right. Note the presence of serpentinised peridotite (s) on top and in the middle right, and the angular to subrounded shape of the quartz detritus (q)on the bottom right and middle left. Crossed polars, 40X.

which are subhedral and prismatic or tabular in shape with size grades up to 1.5 mm long. The pyroxene includes diopside, augite and hypersthene. Magnetite is usually present in small amounts. Serpentinite fragments occur in small size (less than 3 cm long), and consist predominantly of flaky antigorite.

The peridotite fragments are holocrystalline and hypidiomorphic granular and contain olivine which is largely altered, with subsidiary hypersthene, enstatite and magnetite. The peridotites show mesh-structure due to alteration of olivine to serpentine.

The metagabbro fragments are heteroblastic with plagioclase and pyroxene, most of which show parallel orientation. Plagioclase and pyroxene show subidioblastic and nematoblastic textures with size grades up to 3 mm long. The plagioclase consists of labradorite (An 52-68) and shows polysynthetic twinning. The pyroxene includes augite and hypersthene. Hornblende is present in very small amount.

The limestone fragments consist of packstones and grainstones which are composed largely of skeletal grains of benthic foraminifera and calcispheres of planktonic foraminifera, with size grades up to 4 mm long and may constitute up to 60% of the rocks (e.g. 83 TO 18). Detrital grains of calcite of silt size are present in small amounts (less than 5%). Most of the calcispheres are micritised, and some are filled by sparry calcite. The matrix consists of mixed lime mud and micrite. The marlstone fragments consist of wackestone and mudstone containing numerous calcispheres of planktonic foraminifera. The limestone and marlstone fragments are compositionally and texturally similar to those of the Palaeogene Salodik Limestones.

Reddish brown limestone fragments consist largely of micritised calcispheres of microfossils (50%), and minor calcite detritus, set a in matrix of mixed lime mud, micrite and iron oxides giving rise to the reddish-brown colour of the rocks. Texturally and compositionally, the reddish-brown limestone fragments are quite similar to that of the Late Jurassic Sinsidik Beds.

The chert fragments are typically angular in shape, and consist essentially of cryptocrystalline quartz with numerous ghosts of radiolaria. The chert may be derived from the pelagic cover of the ophiolite suite. The calcilutite fragments consist of wackestone and mudstone, most of which contain scattered ghosts of microfossils. The calcilutite may be derived from either the Late Cretaceous Luok Beds or from the Cretaceous pelagic cover of the ophiolite suite.

The coarse clastic facies contains a matrix of mixed coarse grained sandstones, mud, clay and iron oxides, which is compositionally and texturally similar to the fine-grained clastic lithofacies described below. The matrix contains significant amounts of quartz detritus, which is predominantly monocrystalline, up to 2 mm across, mostly subangular in shape (e.g. 83TO19). A calcareous sandy-matrix is also present locally in the coarse clastic facies.

(ii) <u>Fine-grained clastic lithofacies</u>

This facies consists largely of arenites and minor silty shale intercalations, some of which are calcareous. In outcrops they are well bedded, with thickness ranging from 50 cm to nearly 4 metres. Sedimentary features abscribed to the Bouma's (1962) sequence are present in some beds of the arenites, but most of them are incomplete sequences, including truncated and base-cut-out and truncated sequences. The grain size ranges from silty shale to pebbly arenite. Most of the rocks are compact or lithified and hard, and grey to darker grey in colour. Some of the silty shales are calcareous, and contain scarce planktonic and benthic foraminifera.

In thin section, the arenites are grey, yellowish and brownish in colour. They are typically lithic arenites consisting largely of rock fragments and subsidiary mineral detritus and opaque minerals. The lithic fragments include serpentinised peridotites, gabbro, dolerite, basalt, diorite, metamorphics, epidiorite, limestone, marlstone, mudstone and chert, and may constitute up to 65% of the rocks (e.g.83 TO 91A). Serpentinised peridotites predominate among the ophiolite-derived fragments, and form up to 20% of the rocks. They grade up to 4 mm in size. The peridotites are invariably serpentinised, and often show mesh or honeycomb structures due to alteration of olivine. Subsidiary to olivine, the peridotites contain pyroxene which is partially altered and replaced by chlorite and iron oxides. Iron ores are present in small amounts (less than 3%), and are probably of magnetite.

The limestone fragments are subrounded to rounded in shape with size grades up to 2 mm long, and consist of grainstones and wackestones, and are compositionally and texturally identical to those fragments of coarse clastic rocks described previously. These carbonate fragments are similar to the Palaeogene Salodik Limestones.

Chert fragments are typically angular in shape, with size grades up to 1 mm long. Chert fragments may be present up to 15% of the rocks (e.g. 83 TO 91A, 83 TO 93). Few grains of siliceous mudstone of subrounded and tabular shape up to 1.5 mm long are also present.

Mineral detritus includes quartz, feldspar, pyroxene, amphibole and calcite (Plate 4.4A). The quartz grains are mostly monocrystalline, with size up to 0.5 mm long, subangular to angular in shape. Some of them show undulatory extinction, and strained features due to plastic deformation. Cryptocrystalline quartz occurs as smaller grains and in the matrix. Some of them may have been derived from chert and/or recrystallised siliceous mudstone. The feldspar grains consist of both K-feldspar and plagioclase, and are subhedral to subrounded in shape, with size grades up to 2 mm long. Some of the feldspar grains are marginally or zonally altered and replaced by carbonates and sericite. The K-feldspar includes orthoclase and microcline showing a typical cross-hatched texture. The plagioclase consists of albite and oligoclase (An 4-10) and andesine-labradorite (An 34-48) characterised by carlsbald-albite twinning. Some of the grains show a compositional zonation. The occurrence of two types of feldspars suggest two distinctive provenances; K-feldspar and the sodic-plagioclase were ultimately derived from an acidic plutonic associated with volcanic terrains, while the calcic plagioclase grains were derived from the ophiolite suite.

Ortho- and clino-pyroxene are both present as detrital grains. They include augite, hypersthene and diopside, which are partially altered and replaced by chlorite and iron oxides. They usually occur in subhedral and prismatic shape with size grading up to 0.5 mm long, and may be present up to 5% of the rocks (e.g. 83 TO 2). The smaller grains occur in the matrix.

Amphibole detritus occurs as prismatic grains up to 0.5 mm long, and may constitute up to 5% of the rocks. Some of the grains show a wavy or undulatory extinction, which is probably due to plastic deformation. Some of them are brownish in colour and typically highly pleochroic.

Calcite detritus occurs in small amounts (3%), and is subangular in shape, with size grades up to 0.5 mm across. Some of the calcite grains were micritised. Sparry calcite may occur locally as cement.

These fragments are set in a matrix of mixed clay mud, cryptocrystalline quartz and iron oxides. The arenites are texturally and compositionally immature, suggesting very rapid deposition of the sediments.

The silty shales are texturally identical to the arenites, but compositionally some of the silty shales are marlstone consisting of wackestone which contains scarce calcispheres of planktonic foraminifera and minor detritus grains of quartz, feldspar, muscovite, chlorite and opaque minerals set in a lime mud matrix. Most of the calcispheres are micritised and/or infilled by sparry calcite.

D. Biostratigraphy

The Biak Conglomerates very rarely contain fossils. Nannoplankton occurring in the marlstone were identified by Mrs. C. Muller (written comm.) as Gephyrocapsa sp. of Pliocene age. Planktonic foraminifera occurring in marly siltstone and marlstone includes Globigerinoides trilobus (Reuss),Globigerinoides immaturus Le Roy, Orbulina universa D Orbigny and Rotalia sp.,indicating an age younger than Late Miocene (Purnamaningsih, GRDC, personal comm.).

Coralline reefs overlying unconformably the Biak Conglomerates contain scarce planktonic foraminifera including Globorotalia tosaensis Takayani & Saito, Globorotalia tumida (Brady), Globorotalia truncatulinoides D' Orbigny, Pulleniatina obliquiloculata (Parker & Jones) of Pleistocene (N22) age (Budiman,GRDC, personal comm.) Hence, the age of Biak Conglomerates must be older than N22 (Pleistocene).

On the basis of these faunal assemblages the age of the Biak Conglomerates is very late Miocene to Pliocene.

E. Stratigraphic relationship between lithofacies

Outcrops of the Biak Conglomerates indicate that coarse clastic rocks predominate in the succession. Arenites with minor silty shale and calcareous shale intercalations are present in much less amount. The ratio af sand to shale is very high (exceeds 5:1) indicating the very proximal depositional setting of the rocks.

The outcrops also show a typical thinning upwards sequence in the lower part, but a thickening and coarsening upwards sequence in the upper part of the sequence. Conglomerates usually occur at the base of each succession, and transititionally grade up into the overlying pebbly and/or very coarse grained arenite beds. Some of the conglomerates may be associated with silty shale beds. Some of the fine grained clastic rocks show typical features of classical turbidites, but are mostly incomplete sequences. The repetition of alternating beds of conglomerates with sandstone with or without silty shale intercalations suggests a very rapid rate of deposition for the Biak Conglomerates.

F. Discussion and Interpretation

The Biak Conglomerates genetically and compositionally are classified as post-orogenic clastic sediments deposited on top of the collision complex, or more specifically as molasse type sediments. Sedimentologically, the upper part of the succession shows a coarsening and thickening upwards depositional pattern, a very high ratio of sand to shale, with conglomerate exceeding sandstone. While in the lower part the succession shows thinning and fining upwards. The geometry of the exposure forms an U or V shape body, that slightly widens towards the sea. These features may suggest that the Biak Conglomerates may have been deposited in a north-south trending submarine canyon or feeder channel depositional setting. This is also suggested by the southerly directed sedimentation transport of the succession.

The rocks contain very scarce fossils, but the occurrence of nannoplankton and planktonic foraminifera in marlstone intercalations strongly suggests an open marine environment. The upper part of the Biak Conglomerates are interfingered with the Plio-Pleistocene coralline reefs, which also points to a marine depositional setting.

Mega-crossbedding and braided stream features, which are typical of coastal deposits are absent in this succession. However, in places, particularly in the western part of the east arm and Central Sulawesi, a similar succession is interpreted, at least partly, to have been deposited in a terrestrial environments. (Simandjuntak et al.,1982).

The depositional setting of the Biak Conglomerates is largely a marine environment, which became a coastal, river-mouth or deltaic setting during the deposition of the uppermost unit of the succession. The occurrence of subrounded to rounded of clasts, especially clasts of mafic and ultramafic rocks, suggests a long period of transportation of river deposits. This feature implies that the East Arm of Sulawesi was originally in the form of much larger land-mass than its present configuration, providing much longer stream courses, in which these clasts were transported. The Biak river valley appears to have been developed as a feeder canyon (channel) connecting the Tomori Basin with the palaeo-stream courses. The occurrence of both planktonic and benthic foraminifera in the silty shale and arenite beds, indicates that the Biak Conglomerates were deposited in open marine environment. This fact argues that the Biak Conglomerates can not have originated as terrestrial river deposits.

Physiographically two depositional settings may be distinguished : marine deposits mostly occurring near the present coast of the east arm, while subaerial deposits occur in isolated basins on land.

Some of the basins may have been developed as an isolated-graben type basins during and subsequent to the uplift of the collision complex, filled partly in subaerial conditions by coarse clastic rocks (Fig. 4.1).

Compositionally and tectonostratigraphically the Biak Conglomerates reflect the geology of the collision complex, from which the rocks were derived, and on top of which the succession was deposited. The rock fragments and detrital grains can be regrouped into two distinctive suites. They are ophiolite suite and continental margin sequences. The quartz, K-feldspar and muscovite detritus are typically recycled detrital grains and derived from continental margin sediments, i.e. the Jurassic Kapali Beds. The sedimentary fragments, especially the limestone and marlstone are clearly derived from the Palaeogene Salodik Limestones. The reddish-brown limestone fragments are typical of the Late Jurassic Sinsidik Beds.

The ultrabasic and basic and metamorphic fragments have no doubt come from the ophiolite suite. The radiolarian chert fragments are derived from the pelagic cover of the ophiolite suite.

The composition of the Biak Conglomerates also points to the post collision (orogenic) deposition setting, giving rise to the development of molasse type sediments. Similar depositional setting of molasse sediments is fully described and documented elsewhere in the world, such as that Oligocene to Miocene Alpine Molasse, which consists of marine turbidites and terrestrial deposits (Fuchtbauer, 1967), and that the Appalachians, Devonian Catskill facies (Walker and Harms, 1971).

In the East Arm of Sulawesi, the Middle Miocene collision of the Eastern Sulawesi Ophiolite Belt against the Banggai-Sula Platform gave rise to the imbrication and uplifting of the collision complex bringing together the ophiolite and continental sources.

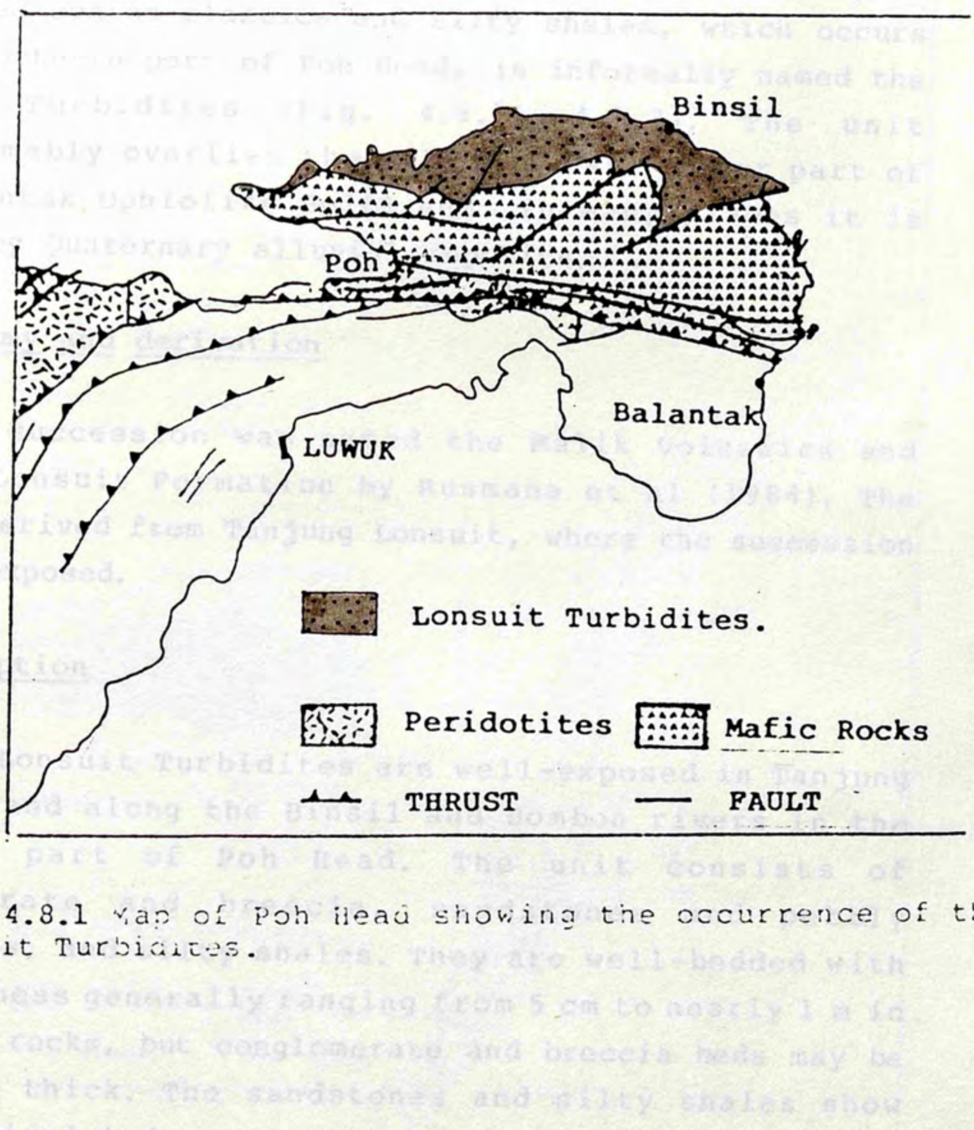


Fig. 4.8.1 Map of Poh Head showing the occurrence of the Lonsuit Turbidites.

4.3.3 LONSUIT TURBIDITES

A. Definition

A volcanigenic sedimentary succession consisting of alternated coarse clastics and silty shales, which occurs in the northern part of Poh Head, is informally named the Lonsuit Turbidites (Fig. 4.8.1; 4.8.2). The unit unconformably overlies the pillow lavas (upper part of the Balantak Ophiolite Belt) and in many places it is covered by Quaternary alluvial deposits.

B. Synonymy and derivation

The succession was named the Malik Volcanics and Tanjung Lonsuit Formation by Rusmana et al (1984). The name is derived from Tanjung Lonsuit, where the succession is best exposed.

C. Description

The Lonsuit Turbidites are well-exposed in Tanjung Lonsuit, and along the Binsil and Bombon rivers in the northern part of Poh Head. The unit consists of conglomerate and breccia, sandstones and pebbly sandstones, and silty shales. They are well-bedded with bed thickness generally ranging from 5 cm to nearly 1 m in the finer rocks, but conglomerate and breccia beds may be up to 7 m thick. The sandstones and silty shales show parallel-sided beds.

In outcrop the sandstones show a regular pattern with a succession consisting of pebbly sandstone in the lower part grading through very coarse to fine grained sandstones to the top of the sequence. This pattern can be developed in a single bed (i.e. a turbidite bed) and in a

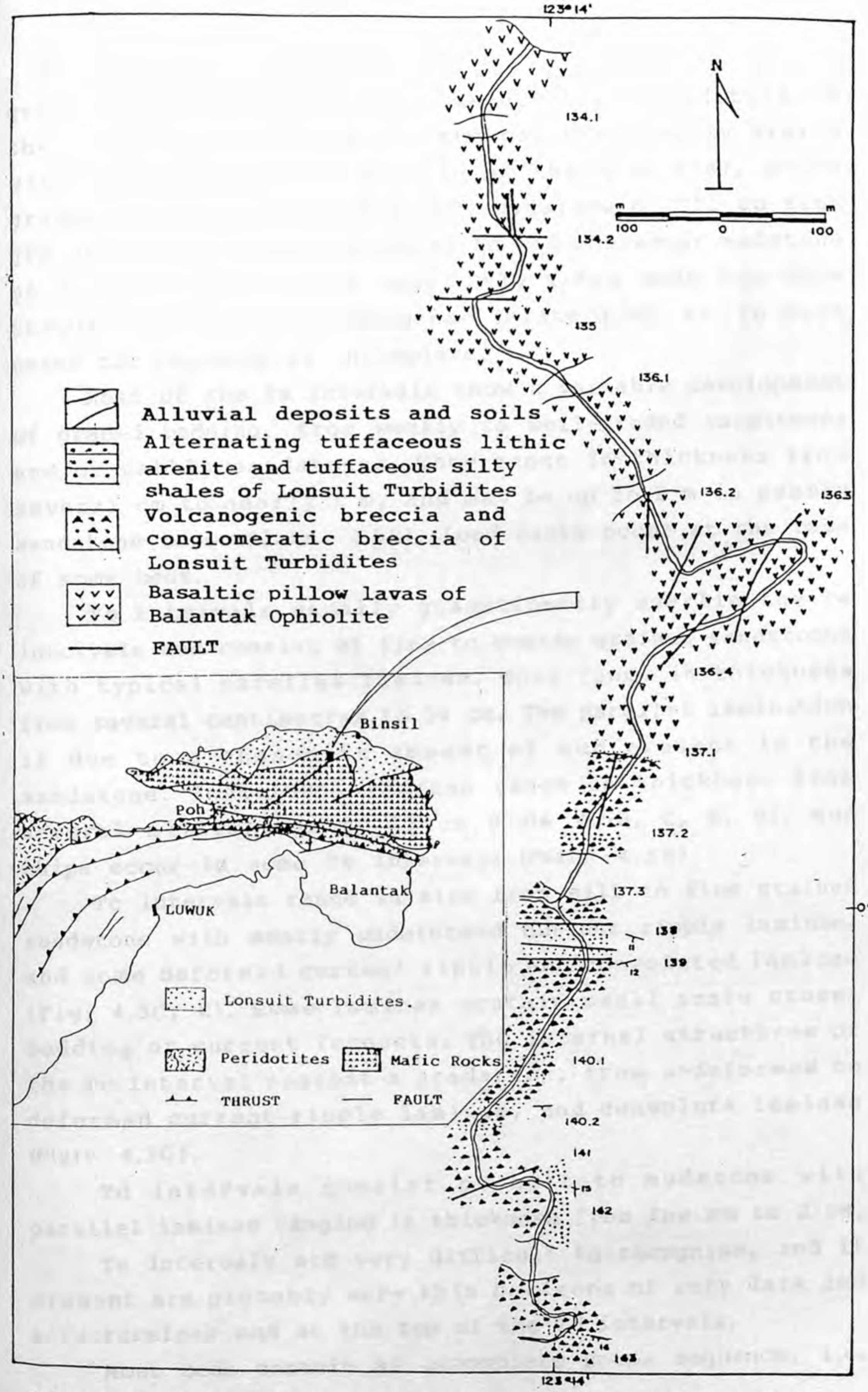


Fig. 4. 8 .2 Geological traverse map of Bombon River, showing the occurrence of the Lonsuit Turbiites.

group of turbidite beds (i.e. megacyclic turbidites). In the case of turbidite beds, the pattern usually starts with coarse to pebbly arenite at the base (Ta), which grades through medium grained sandstone of Tb, to fine grained of Tc, to siltstone of Td and shales or mudstone of Te at the top of the beds. Only a few beds may show complete Ta-e Bouma's sequence (Plate 4.5C, F). In most cases the sequence is incomplete.

Most of the Ta intervals show a variable development of graded bedding, from weakly to well-graded sandstones and/or pebbly sandstones. They range in thickness from several cm to nearly 1 m, and may be up to 2 m in pebbly sandstone beds (Plate. 4.5E). Load casts occur at the base of some beds.

Tb intervals usually gradationally overlie the Ta intervals and consist of fine to coarse grained sandstones with typical parallel laminae. They range in thickness from several centimetres to 50 cm. The parallel lamination is due to a change in amount of mud present in the sandstone. Individual laminae range in thickness from several millimetres to 15 cm (Plate 4.5B, C, D, E). Mud chips occur in some Tb intervals (Plate 4.5E)

Tc intervals range in size from silt to fine grained sandstone with mostly undeformed current ripple laminae, and some deformed current ripple and convoluted laminae (Fig. 4.5C, E). Some laminae contain small scale cross-bedding or current foresets. The internal structures of the Tc interval suggest a gradation, from undeformed to deformed current ripple laminae, and convolute laminae (Plate 4.5C).

Td intervals consist of silt to mudstone with parallel laminae ranging in thickness from few mm to 2 cm.

Te intervals are very difficult to recognise, and if present are probably very thin horizons of very dark and structureless mud at the top of the Td intervals.

Most beds contain an incomplete Bouma sequence, i.e.

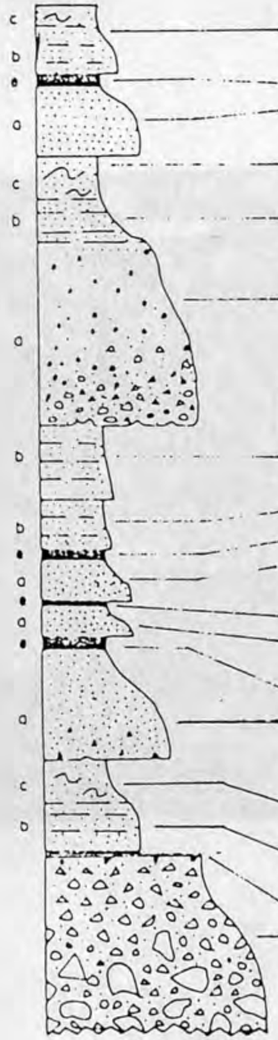


Fig. 4.12A, B Detailed section of parts of the Lonsuit Turbidites occurring in Bomcon River (d3TO139).

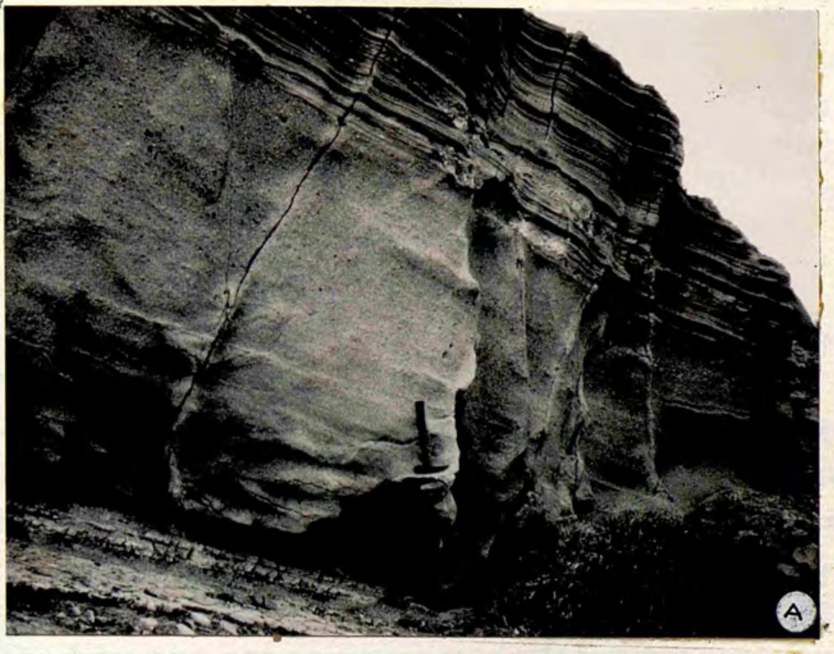


Plate 4.5A Photograph of the Lonsuit Turbidites on steep cliff of Tanjung Lonsuit (83 TO 88), showing a very thick bed of pebbly lithic arenite, alternating with thinly bedded turbidite deposits at the top and bottom of the sequence.

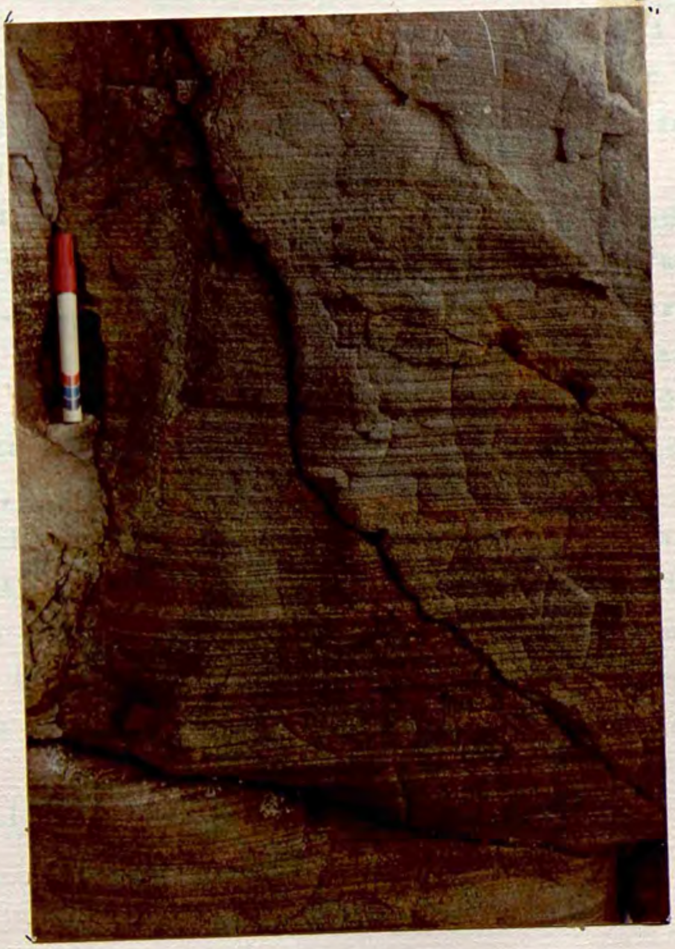


Plate 4.5B Photograph of exposure of Lonsuit Turbidites on coast of Tanjung Lonsuit (83 TO 88.3), showing thin parallel laminae in coarse grained tuffaceous lithic arenites.

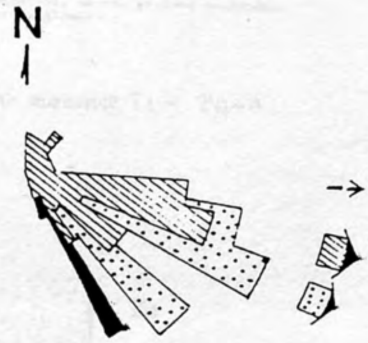
truncated, base-cut-out and combination of these two. Ta-d is the commonest sequence in the thicker beds, and Td-e in the thinner beds. Ta is rarely present as single bed, and if present usually forms amalgamated beds, in which two or more beds are welded together with very thin and usually discontinuous intercalations of mudstone.

The coarser clastic rocks, including conglomerate, breccia and pebbly conglomerate, usually occur as lensoid and/or channel fill beds. Some of the conglomerate and pebbly conglomerate beds show grading, indicated by a decrease in both size and amount (volume) of clasts toward top of the beds. In some beds the clasts may show fabric alignment, i.e. the long axis is imbricated up stream as indicated by the other current indicators, such as small scale current foresets and flute casts (Plate 4.5E). Some of the thicker beds of conglomerate and pebbly conglomerate may show an internal stratification 2 cm thick or more, indicated by the occurrence of a clast-supported stratum, which is contrasted with the alternating sand-supported strata. The thicker strata may attain thicknesses up to 30 cm. These rocks were deposited by either grain flows or liquidised flows.

Detailed sections made in Tanjung Lonsuit and the Bombon river (Fig. 4.12B; 4.12A) show megacyclic turbidites. Each megacycle consists of several beds of turbidite, which are repeated. The lower part of each megacycle is usually formed by lensoid breccia, conglomerate or pebbly sandstone beds, which towards the upper part, pass into sandstones interbedded with silty shales. Each megacyclic turbidite shows a fining-upward facies pattern in both grain size and bed thickness. In Fig. 4.12A is shown the development of at least 8 megacyclic turbidites in continuous exposure about 125 m long in the Bombon river (Localities 83 TO 139 to 83 TO 140).

Overall the Lonsuit Turbidites show a very high sand-

Fig. 4.10 Paleocurrents indicate a northwest source area of the Lonsuit Turbidites



27 readings on the long axis (B-axis) imbricated component of conglomerates

20 readings on current foreset in Tc interval

6 readings on fold axis of convolute lamination in Tc interval

Only 2 readings on flute cast which is poorly developed on the base of arenite beds.

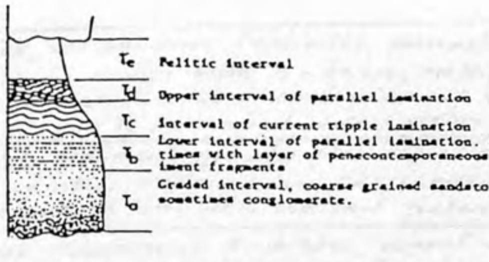


Fig. 4.11.1 Complete sequence T_1 = Ta–e

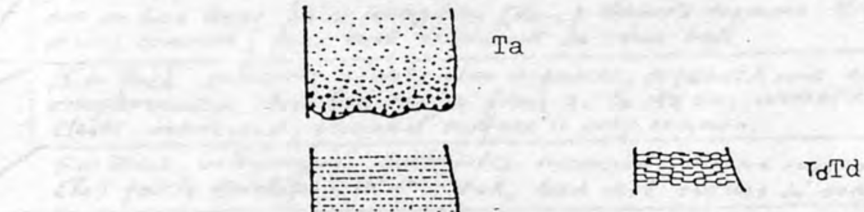


Fig. 4.11.2 Configuration of incomplete sequence, base cut-out units (A), truncated units (B), and truncated base cut-out units (C).

Fig. 4.12A Diagram showing the occurrence of at least 8 megacycles turbidites in the Bombon river section (83 TO 139 - 140.2). Each megacycle consists of very coarse clastics (sediment gravity flow deposits) in the lower part and a succession of turbidite in the upper part.

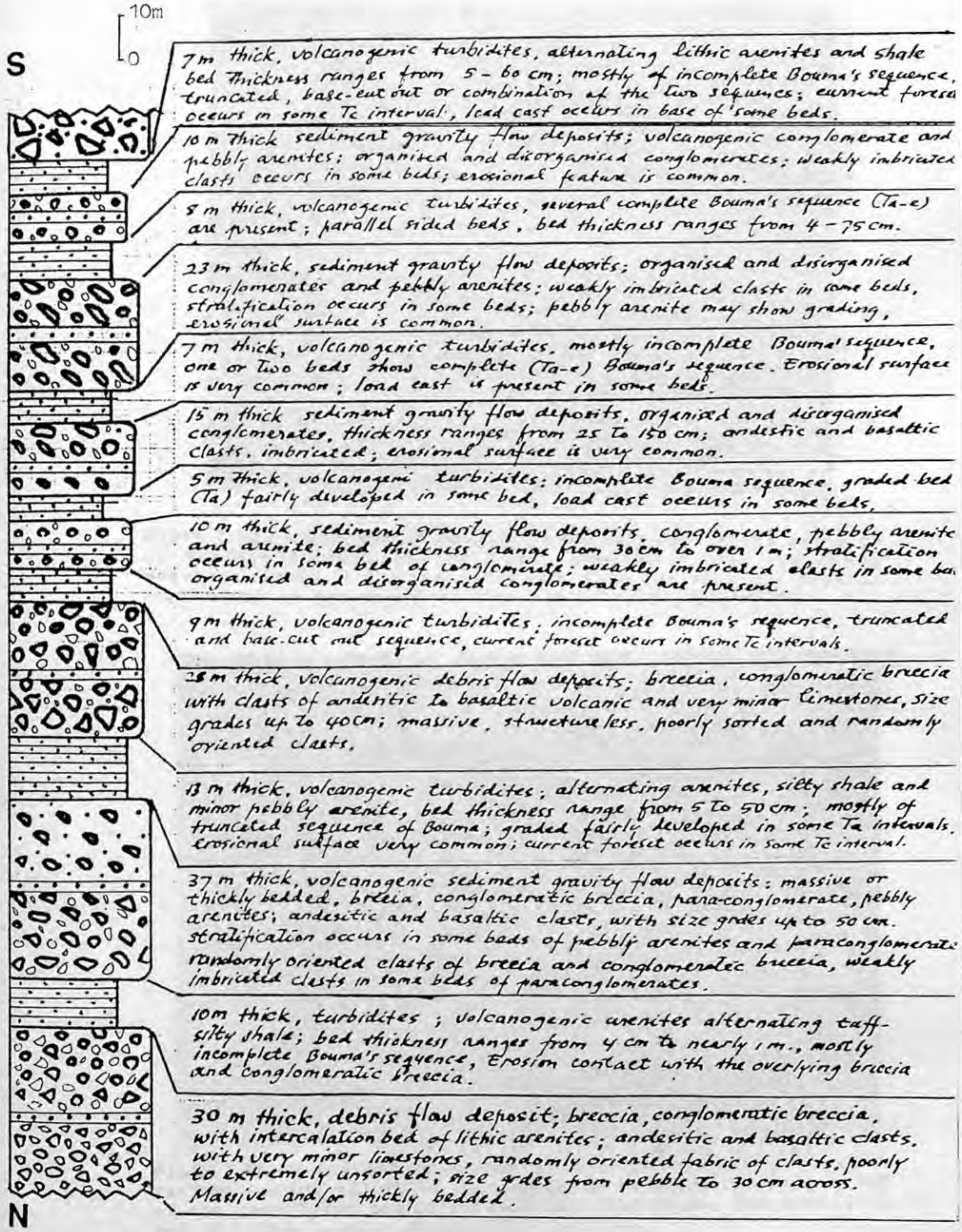


Plate 4.5C Photograph of outcrop of Lonsuit Turbidites on coast of Tanjung Lonsuit (83 TO 88.1), showing a complete Bouma sequence (Ta-e) in the turbidite bed. Note the volcanic pebbles (equivalent to mud chips) at the base of weakly graded Ta.

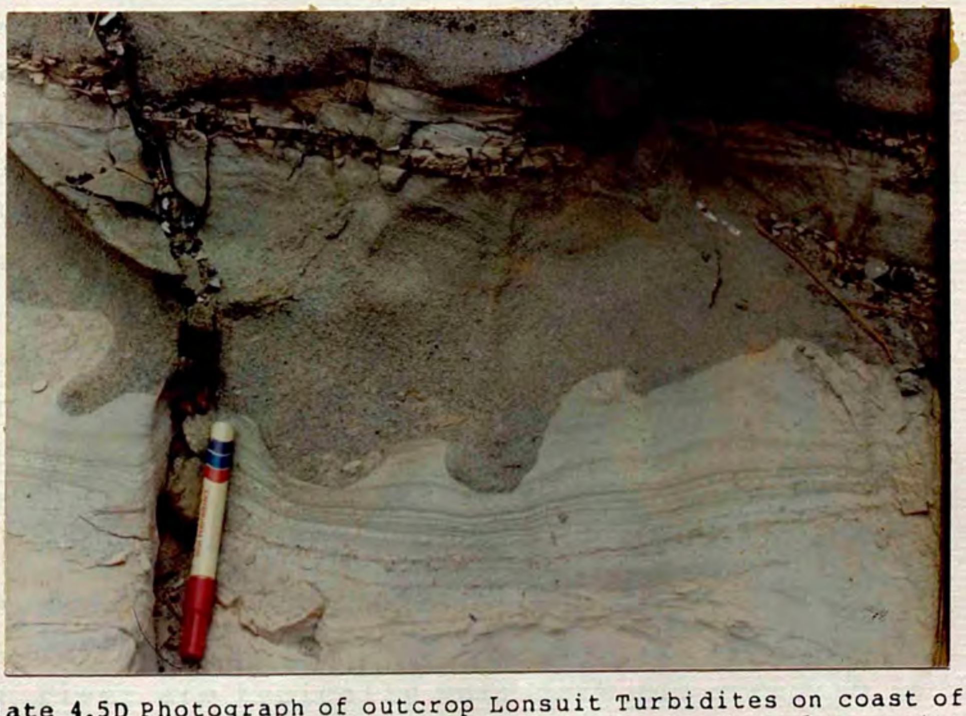


Plate 4.5D Photograph of outcrop Lonsuit Turbidites on coast of Tanjung Lonsuit (83 TO 88.2), showing an erosional surface on top of the underlying bed of tuffaceous silty shale. The overlying bed of lithic arenite shows a weak grading.

shale ratio = 8:1 (Lovell, 1970), high proximality index, Pl = 70% (Walker, 1967) and a high proportion of coarse clastics, indicating that the succession was deposited in a proximal depositional setting. The distal part of the succession is absent.

The ratio of sand to shale is defined on the basis of proportions of individual beds of sand to those of shale (Lovell, 1970). The proximality index (Pl) of Walker (1967) is defined by using the equation Pl = A - (A-E) + 1/2B, where A and B are the percentages of beds in the group of turbidites starting with division A and B respectively, (A-) is the percentages of beds in the group which are thin (usually less than 3 cm thick), fine grained with no internal structures and grade smoothly up to the mudstone or shale. These two values of depositional indicators enable one to obtain a result which is representative of formation as whole from only a small part of the succession.

Sedimentological features of the Lonsuit Turbidites indicate that the clast-supported conglomerates and breccias were deposited by the mass-gravity movement of debris flows, and the alternating sandstone and silty shale sequences are classical turbidites. The thicker and structureless sandstone beds were deposited by grain flows and/or fluidized sediment flows. These depositiononal mechanisms repeatedly operated, resulted in the development of megacyclic turbidites.

The sediments are gently folded with fold axes trending in east-west direction. Fractures and faults are frequently observed within the succession.

The rocks exposed in Tanjung Lonsuit are commonly dark-grey in colour, and consist largely of alternating sandstones and silty shales with lesser amounts of conglomerate and breccia. In contrast, rocks exposed in the Bombon river are typically very dark and greenish in colour due to presence of abundant mafic detritus,

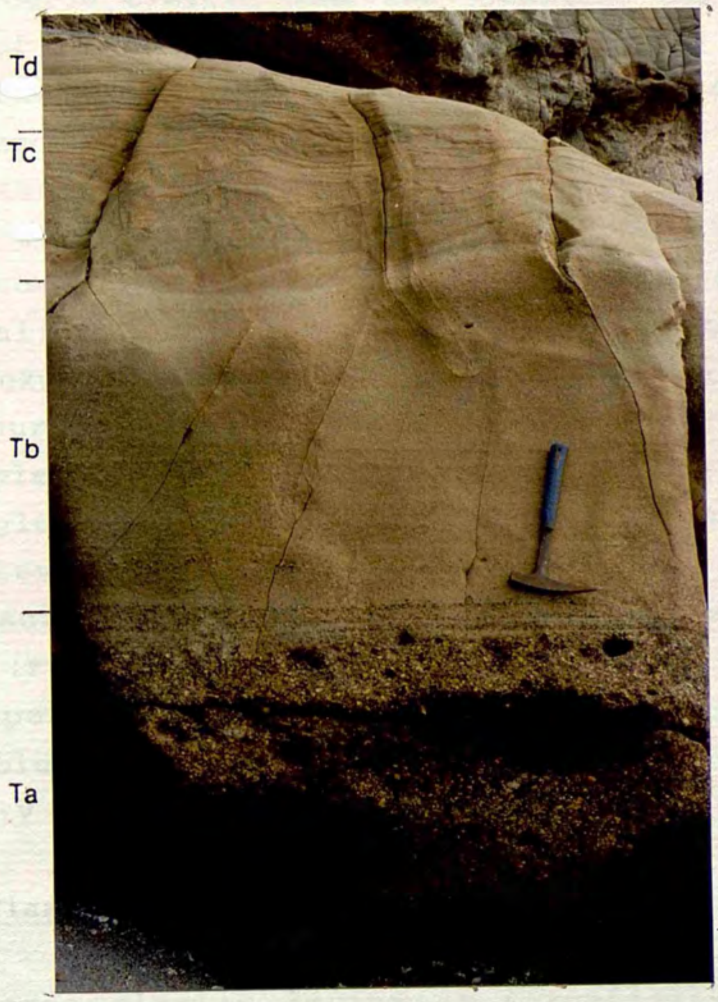


Plate 4.5E Photograph of very thick (nearly 2 m) turbidite bed of Lonsuit Turbidites exposed on coast of Tanjung Lonsuit (83 TO 88.3), showing the Ta interval at bottom is composed of poorly graded pebbly arenite. Note the uppermost parallel laminae of Td the overlain by very thin and structureless tuffaceous shale.

volcanic glass and iron oxides. The rocks consist of equal amounts of coarse clastic beds and classical turbidites, consisting of alternating sandstones and silty shales. Megacyclic turbidites are much better developed in the Bombon section.

Due to the limited sections which have been studied in detail, the thickness of the succession is not definitely known. However, based on the cross-section made along the Malik river, Rusmana et al. (1984) estimated that the thickness of the unit is at least 100 metres.

Palaeocurrent measurements on current foreset or current ripple lamination and imbricated fabric of long axis of conglomerate clasts, and flute casts which are poorly preserved on the base of sandstone beds, suggest that the Lonsuit Turbidites were derived from a northwest source area (Fig. 4.10).

For purposes of description and interpretation, the Lonsuit Turbidites are divided into : (i) coarse clastic and (ii) silty shale or mudstone lithofacies.

(i) Coarse Clastic Lithofacies.

The coarse clastic facies is essentially clast-supported sediments, including sandstone, pebbly sandstone, conglomerate and breccia. In outcrops these rocks are thickly bedded, or massive with thickness ranging from a few centimetres in the arenites to 6 metres in the breccia or conglomerate. The outcrops suggest that the conglomerate and breccia occur as lensoid and/or channel fill deposits with the basal contact of each bed always showing erosional features (Plate 4.5D, E).

The clasts are predominately composed of basaltic and andesitic volcanic rocks; a few sandstone fragments may be present locally; with size grades from granule up to 50 cm across. Fragments nearly 1 m long may be present in the breccia. Fabric orientation ranges from randomly oriented clasts in the breccia and some of conglomerates, to weakly imbricated clasts in some beds of conglomerate. The framework of clasts of pebbly conglomerate range from very poorly sorted conglomeratic breccia and disorganised conglomerates, to moderately sorted organised conglomerates. The clast shape also varies from extremely angular fragments of breccia, to the tabular or elongated and subrounded clasts in the conglomerates and pebbly conglomerates.

In thin section the coarse clastic rocks contain a matrix consisting of mixed clay, volcanic glass, pyroxene, chlorite and iron oxides darker, grey or greenish in colour. The matrix may constitute 30% of the organised conglomerate and up to 70% in the paraconglomerate or pebbly conglomerate.

Most of the basaltic clasts are porphyritic, but some are hyalopilitic in texture. Plagioclase may form up to 40% of the clasts (e.g. 83 TO 140.1), and consists of labradorite (An45-48). Plagioclase occurs as phenocrysts and as microlites in a glassy groundmass, which shows hyalopilitic texture. The plagioclase phenocrysts are commonly subhedral, some are prismatic or tabular in shape, with size grades up to 2.2 mm long. Some of the plagioclase may show typical compositional zoning (e.g. 83 TO 140.1). Some of the phenocrysts contain tiny inclusions of glass, pyroxene and iron ores.

The groundmass of the andesitic clasts is usually fine grained, and in some of them is almost entirely a dark brown to opaque glass (e.g. 83 TO 139.2). Less glassy andesitic clasts contain a groundmass with plagioclase laths forming phenocrysts and granules of augite (e.g. 83 TO 141). Some of the plagioclase phenocrysts are partly altered and replaced by clay, carbonate, sericite and zeolite. Marginal alteration is common, but zonal alteration occurs in some crystals (e.g. 83 TO 143). The bulk of the plagioclase is fresh and

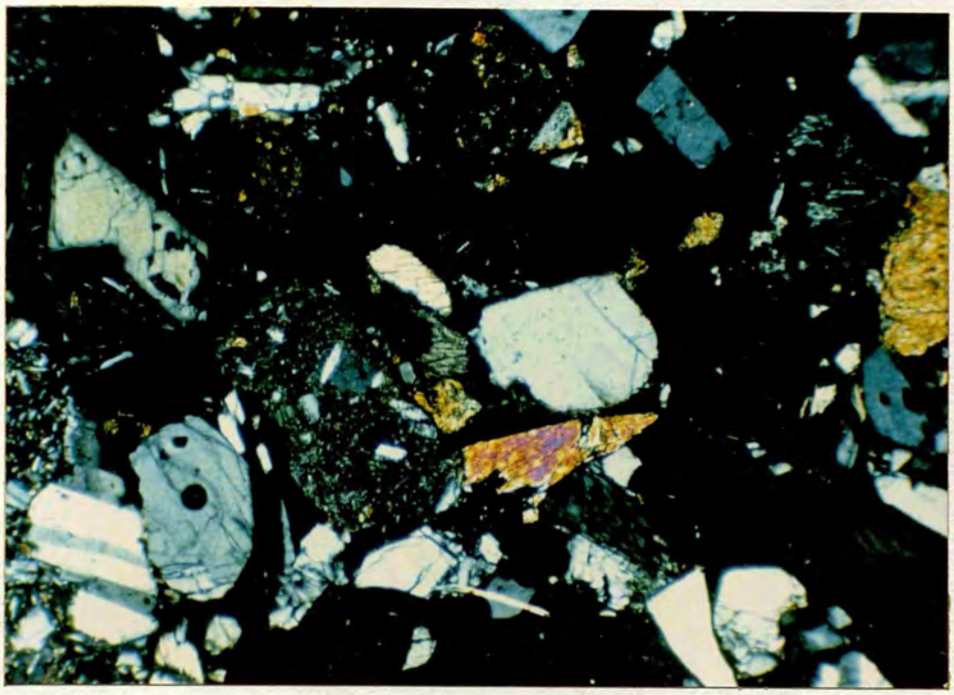


Plate 4.6A Photomicrograph of volcanogenic lithic arenite of Lonsuit Turbidites (83 TO 88.1), showing subhedral and angular to subangular grains of plagioclase detritus. Angular lithic fragments of basaltic and andesitic volcanics and prismatic grains of pyroxene are also present. Crossed polars, 40X.

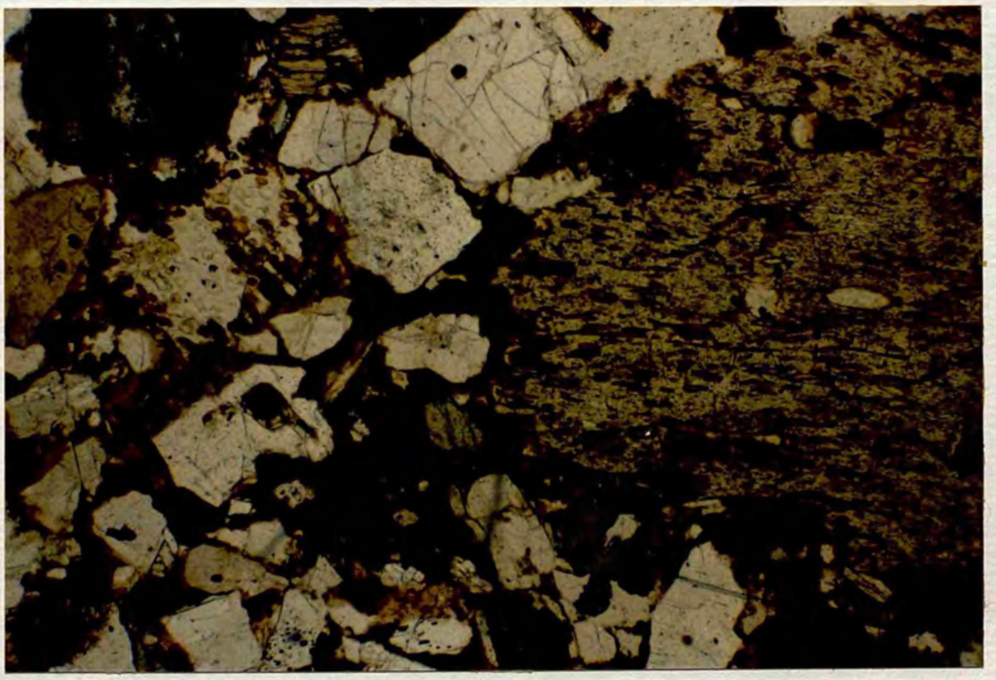


Plate 4.6B Photomicrograph of volcanogenic arenite of Lonsuit Turbidites (83 TO 88.3) showing the occurrence of large fragments of andesitic tuff (right) and angular grains of plagioclase detritus. Plane polarised light, 40X.

angular and elongated/prismatic in shape. Many of the crystals show marginal intergrowth with the silty clay matrix.

Pyroxene occurs as phenocrysts and groundmass and may form up to 15% of individual clasts (e.g. 83 TO 143). There was probably much more pyroxene present originally, but some has been altered to chlorite, uralite and iron oxides. The pyroxene is augite, which is anhedral with a short-prismatic shape, and grades up to 1 mm long (e.g. 83 TO 143). The pyroxene phenocrysts include greenish diopsidic augite and hypersthene with distinctive red to pale green pleochroism; the former being usually more abundant. Simple and multiple twinning are frequently developed in the pyroxene phenocrysts. Some of the pyroxene phenocrysts are interlocked with plagioclase phenocrysts. Iron oxide inclusions occur in some pyroxene phenocrysts. In some rocks the pyroxene is almost completely altered to chlorite, uralite and iron oxides (e.g. 83 TO 139.2).

Zeolite occurs as an alteration product of plagioclase in both phenocysts and groundmass. It is commonly anhedral and shows fan or spherulitic textures. Some of the zeolite also occurs as vesicle fillings, forming amygdales. Zeolite may make up to 15% of the rocks (e.g. 83 TO 143).

Iron oxides occur as very fine opaque euhedra scattered throughout the rocks. Iron oxide also occurs in amygdales, forming up to 3% of the rock in 83 TO 140.2. Some of the opaque grains are concentrated in localised areas, probably derived from the alteration of pyroxene.

Chlorite is also commonly present as a product of the hydrothermal alteration of pyroxene, and may constitute up to 15% of the rock (e.g. 83 TO 143). Some of the chlorite occurs filling vesicles. Other additional minerals present in very minor amounts include carbonate, epidote and sericite as an important products of the hydrothermal alteration of the basaltic-andesitic fragments occurring in the coarse clastic rocks of Lonsuit Turbidites.

The sandstone fragments consist of tuffaceous lithic arenites and feldpatholithic arenites, angular to subrounded in shape with size grades up to 30 cm across. Lithic fragments in these clasts consist largely of basaltic and andesitic volcanics similar to those clasts described previously, and minor vitric tuffs. Detrital grains consist largely of angular to subrounded plagioclase, mostly labradorite with subsidiary andesine, pyroxene and iron oxides and very minor quartz. Some of the plagioclase and pyroxene are altered. The quartz detritus are typically anhedral and clear, indicating a volcanic origin.

Sparry calcite is also present as cement and filling pore-space, suggesting a secondary origin by alteration of plagioclase (e.g. 83 TO 88.3).

(ii) Silty shale Lithofacies

This facies is also characterised by the occurrence of volcanic glass and tuffs and is thinly bedded; the beds being usually less than 15 cm thick. As described previously, these rocks commonly occur in the upper intervals of the turbidite bed. In a complete turbidite bed, the silty shales are usually gradationally underlain by fine grained sandstones. In the case of base-cut-out turbidite beds, the silty shales show sharp contacts with the underlying beds. Erosional surfaces are a common feature occurring on top of the silty shale beds. Shale or mudstone is also present in the so called the 'real turbidite' beds of Walker (1979), that is a thin bed (3 cm thick or less) consisting of fine grained and structureless sandstone which grades smoothly up into mudstone (or shale). Thinly bedded silty shale is also present as an intercalation between two or more thickly bedded sandstones. This is probably related to the development of amalgamated sandstone beds (Walker, 1979).

In thin section, the silty shales are grey, dark grey, greenish and brownish in colour. Compositionally, the silty shales are identical to the coarse clastic rocks and contain volcanic fragments and detrital ores, plagioclase, pyroxene and minor quartz of silt sizes, set in a matrix of altered volcanic glass, clay-mud and iron oxides. In some rocks, the volcanic glass is largely devitrified to clay, but its flow structure is still visible (e.g. 83 TO 88.1). The volcanic glass and clay may constitute up to 80% of the rocks with detrital grains of plagioclase, pyroxene and ores amounting to 20% in a vitric tuffaceous siltstone (e.g. 83 TO 138.2). In some silty shale beds, chlorite-supported laminae alternate with volcanic glass and clay-supported laminae. The laminae are easily observed by the alternation of dark-greenish and dark or grey laminae (e.g. 83 TO 142.1).

Additional grains, probably of secondary origin include prismatic and anhedral apatite of silt size (3%) which is usually associated with carbonate (7%) and small amounts of very fine grained (0.05 mm or less) epidote scattered throughout the rock (e.g. 383 TO 137.2).

The tuffaceous silty shales occurring in the Bombon river contain more abundant chlorite than those in the Tanjung Lonsuit area. But quartz detritus, although in very small amounts (less than 3%) is present in the silty shales of Tanjung Lonsuit (e.g. 83 TO 88.1), and is absent in the silty shales of the Bombon river area.

D. Stratigraphic relationship between lithofacies

The outcrops of Lonsuit Turbidites in both the Bombon river and Tanjung Lonsuit areas indicate that the succession consists of alternating coarse clastic rocks, sandstones and silty shales. The alternating feature is well-documented in the turbidite beds and in the development of megacyclic turbidites as well.

The megacycles, usually comprise coarse clastics including conglomerates and breccias in the lower part, which grade up into a classical turbidite sequence in the upper portion of each megacycle. The clast and matrix-supported breccias and/or conglomerates commonly occur in form of lenses or channel fills. The classical turbidites consist of alternating sandstones and silty shales, and are usually developed as parallel-sided beds.

The occurrence of abundant coarse clastic rocks in the Bombon river area and the abundant classical turbidites in the Tanjung Lonsuit area suggest that the succession is more proximal to the west. The sedimentary transport indicators show that the succession was derived from the northwest. These two features suggest that the distal part of the Lonsuit Turbidites occurs in the area further to the east (i.e. in the North Banda Sea).

E. Biostratigraphy and age determination.

No fossils have been collected from the sediments in Tanjung Lonsuit and Bombon river sections. However, during geological mapping of the East Arm of Sulawesi (Rusmana et al. (1984), planktonic foraminifera, including Globigerinoides sp. and Globorotalia peripheroacuta Blow & Banner, of Late Miocene to Pliocene age were collected from lithic arenites in one locality (82 UH 30A) in the Malik river, just to the east of the Bombon river section.

The outcrops in the Bombon river section (i.e. 83 TO 137.1) also suggest that the Lonsuit Turbidites unconformably overlie the pillow lavas (upper part of the Balantak Ophiolite). Hence, they must be younger than the lavas.

K-Argon dating (Dr. Snelling, British Geological Survey, person. comm.,1985) of basaltic clasts of conglomerate and breccia collected from the Bombon river section and Togian Islands, gave an age of 34.79 +/- 1.06 MY (82 AK 68), 33.95 +/- 1.83 MY (83 TO 141) and 23.07 +/- 3.19 MY (82 AK 57), which range from Early to Late Oligocene. This age is similar to the age of volcanic rocks occurring in the western part of Sulawesi (Sukamto, 1975b). This feature might suggests that the Lonsuit Turbidites are derived from the eroded Oligocene volcanic terrane in the western part of the Sulawesi, and hence, their age must be younger than Oligocene.

These K/Ar ages are compatible with the Late Miocene to Pliocene ages indicated by planktonic foraminifera described above.

F. Discussion and interpretation

The sedimentology of the Lonsuit Turbidites indicates that the succession is typically a product of sediment gravity flows, deposited in a submarine fan.

Compositionally, the succession consists, wholly of volcaniclastic sediments, suggesting a depositional setting in a basin adjacent to an island arc-system produced by the convergence of oceanic plate and oceanic volcanic arc or continental plate, such as trench or forearc basins.

The sediments are texturally and compositionally immature indicating the low kinetic energy of relatively deep depositional environments, as well as a relatively rapid erosion, transport and deposition and direct derivation from volcanoes to site of deposition. Alternatively, although very unlikely, it is possible, that the Lonsuit Turbidites were partly derived from the reworking of Palaeogene volcanic terrane sequences.

The development of megacyclic turbidites appears to be closely related to two main geological factors: (i) the tectonic behaviour of the source area and (ii) the mechanism involved in the transportation and deposition of the rocks.

As the Lonsuit Turbidites are essentially derived from a volcanogenic provenance, volcanic activity is likely to be characteristic feature of the source area, prior to and during the deposition of these rocks. Depositional mechanisms involved appear to have been dominated by the sediment gravity flows, and earthquake activity may have been one of the more important agents initiating or even accelerating these gravity flows. Separate gravity flows were instrumental in the transport and deposition of each bed or cycle within the megacyclic turbidites.

The megacyclic turbidites in the Lonsuit Turbidites consist essentially of coarse clastics and classical turbidites. The development of the megacycles can be explained as follows:

Volcanic activity produces debris ranging in size from volcanic ash to boulder materials. Debris flow is the main mechanism for transporting and depositing the coarse clastic rocks, and turbidity currrents for depositing the classical turbidites. Grain flow and fluidised sediment flow might be also responsible for transporting and depositing some beds of the coarse clastics and some of the thicker and massive sandstone beds. In the case of development of incomplete megacycles, it might be due to the relatively low intensity of volcanic activity producing materials largely of sand to volcanic ash size, with a very minor conglomerate fraction. Alternatively, the volcanic activity may have produced abundant debris ranging in size from ash to boulder, and the occurrence of particular rocks may be result of gravity flow mechanisms in producing deposits whose characteristics and size grade simply reflected their sorting and distance from the source area as well.

It is also quite possible that a high velocity turbulent current may have rapidly overtaken a relatively slow moving debris flows or grain flows (or combination of both) resulting in development of megacyclic turbidites which consist mainly of classical turbidites, such as those occurring in the Tanjung Lonsuit area.

The development of megacyclic turbidites in the Lonsuit Turbidites is essentially due to the change in character of sediment gravity flows during transport. Simandjuntak (1977) described and discussed similar origin and development of megacyclic turbidites in the Cobbadah District, NSW, Australia.

Fisher (1984) reviewed the transport, processes and deposition of submarine volcanic rocks, and observed, that such flow transformation may occur when there is a change from turbulent to laminar flow or vice-versa, when a mass flow segregated by gravity into a concentrated laminar under flow and an overriding dilute turbulent cloud or by mixing with water at the flow margins. Fisher (op.cit.) suggested, further, that turbidity currents may be generated in the fronts of slumps or debris flows (cf. Hampton, 1972) and that pyroclastic flows may pass laterally into lahars and turbidity currents. Wright and Mutti (1981) pointed out, however, that it will often be difficult to distinguish sediment gravity flow deposits generated directly by eruptions, from those which consist of reworked pyroclastics, unless a passage into primary pyroclastic flow deposits can be traced laterally.

Bailes (1980) pointed out, that volcaniclastic turbidites are particularly common in such settings, and may be deposited as submarine fans, supplied from point sources, such as submarine channel mouths. Alternatively, Storey & MacDonald (1984) based on a study on the processes of formation and filling of a Mesozoic back-arc basin on the island of South Georgia, suggested that there may be a linear sediment input from the arc into the basin along the active basin margin, in which case non-fan turbidites develop.

Simandjuntak (1979) described and discussed the depositional setting of Miocene vocaniclastic sediment gravity flow deposits in Pangandaran-Cilacap, SW Java, and their bearing on the tectonic development of the southwestern part of Indonesia.

Mitchell (1970) described and interpreted a variety of submarine volcanic sediment gravity flows from an island-arc setting, in Malekula Island, New Hebrides. Likewise, the Lonsuit Turbidites are believed to be a product of sediment gravity flows in an island arc setting deposited in submarine fan depositional setting.

Paleocurrent indicators suggest that the sediments were derived from source areas to the northwest. The absence of peridotites, gabbro, dolerite, chert and limestone fragments proves that the Lonsuit Turbidites were not derived from the ophiolite or the collision complex, which must therefore have been completely buried in the forearc setting. It is suggested that the Lonsuit

Turbidites were derived from the Late Palaeogene - Neogene volcanic arc of the Western Sulawesi Volcano-Plutonic Belt (WSVB) and were originally situated in the Gorontalo area and has been displaced eastwards for 150 km along the balantak Fault to reach its present position in the East Arm of Sulawesi.

Compositionally, the andesitic-basaltic constituent of the Lonsuit Turbidites corresponds to the Late Palaeogene to Neogene volcanic arcs of the Western Sulawesi Volcano-Plutonic Belt.

4.4 POST-COLLISION STRUCTURES IN THE EAST ARM OF SULAWESI

The most significant structures occurring in the East Arm of Sulawesi subsequent to collision of the BSP and ESOB, are that strike-slip fault, including the Balantak Fault System, Toili Fault, Ampana Fault and Uekuli Fault.

In general the post-collision coarse clastic sediments (i.e. Batui Group) are weakly folded; but, in places, particularly within the Batui thrust, the rocks may be folded more tightly. Joints and fracrures are frequently observed occurring in the Batui Group.

4.4.1 THE BALANTAK FAULT SYSTEM

The Balantak Fault system consists of several faults, all trending in an east-west direction, and extending to Tanjung Api in the west and to the North Banda Sea in the east, up to 85 km long (Fig. 4.13).

Physiographically, the northern part of Poh Head seems to fit the region, off-shore north of Ampana-Bunta and just to the south of the Togian Islands. Lithologically, the southern margin of the northern portion of the Poh Head consists of a narrow belt of serpentinised peridotites with a small exposure of imbricated metamorphic rocks in the Siuna river (Rusmana et al., 1984). A similar rock association occurs in much larger blocks in the Bunta-Ampana-Uekuli area. The northern part of Poh Head is covered by the volcanogenic Lonsuit Turbidites. Similar rocks also occur in the southern part of the Togian Islands.

The exposures in Poh Neck and along the Balantak coast show evidence, in the occurrence of horizontal slickensides, of movement in an east-west direction (Plate 2.9B). Some of the slickensides are superimposed on older slickensides indicating vertical movement.

The horizontal movement along the Balantak Fault

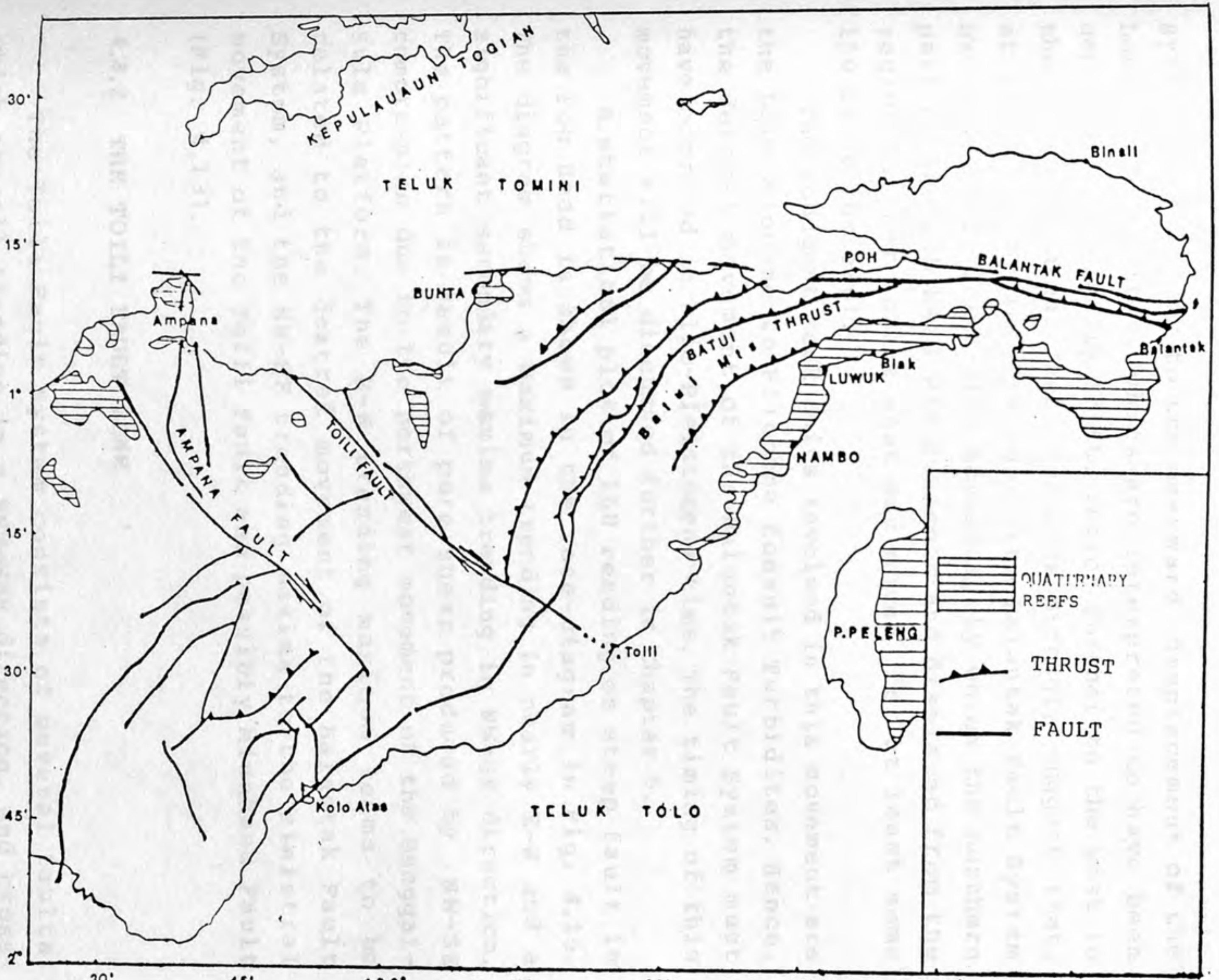


Fig. 4.13 Map showing structural configuration of the East Arm of Sulawesi.

System corresponds to the eastward displacement of the Lonsuit Turbidites, which are interpreted to have been deposited originally in the region further to the west in the Gorontalo area. These features strongly suggest that, at least in the later stages, the Balantak Fault System had a dextral strike-slip movement, by which the northern part of the Poh Head was detached and displaced from the region further to the west and moved for at least some 150 km to the east.

The youngest rock units involved in this movement are the Late Miocene to Pliocene Lonsuit Turbidites. Hence, the dextral movement of the Balantak Fault System must have occurred in Plio-Pleistocene time. The timing of this movement will be discussed further in Chapter 5.

A statistical plot of 160 reading on steep fault in the Poh Head is shown in the rose-diagram in Fig. 4.16. The diagram shows a maximum trending in nearly E-W and a significant secondary maxima trending in NW-SE direction. The pattern is result of pure shear produced by NW-SE compression due to the northwest movement of the Banggai-Sula Platform. The E-W trending maximum seems to be related to the dextral movement of the Balantak Fault System, and the NW-SE trending maxima to the sinistral movement of the Toili Fault and possibly Minahasa Fault (Fig. 4.13).

4.4.2 THE TOILI FAULT ZONE

The Toili Fault System consists of several faults, which are all trending in a NNW-SSE direction, and cross-cut the middle part of the East Arm of Sulawesi. On land the fault system may be up to 90 km long. The north end of the fault dies out in Tomini Bay and the southern end is covered by alluvial deposits and perhaps also dies out in Tolo Bay (Fig. 4.16).

The Toili Fault System is interpreted as a sinistral

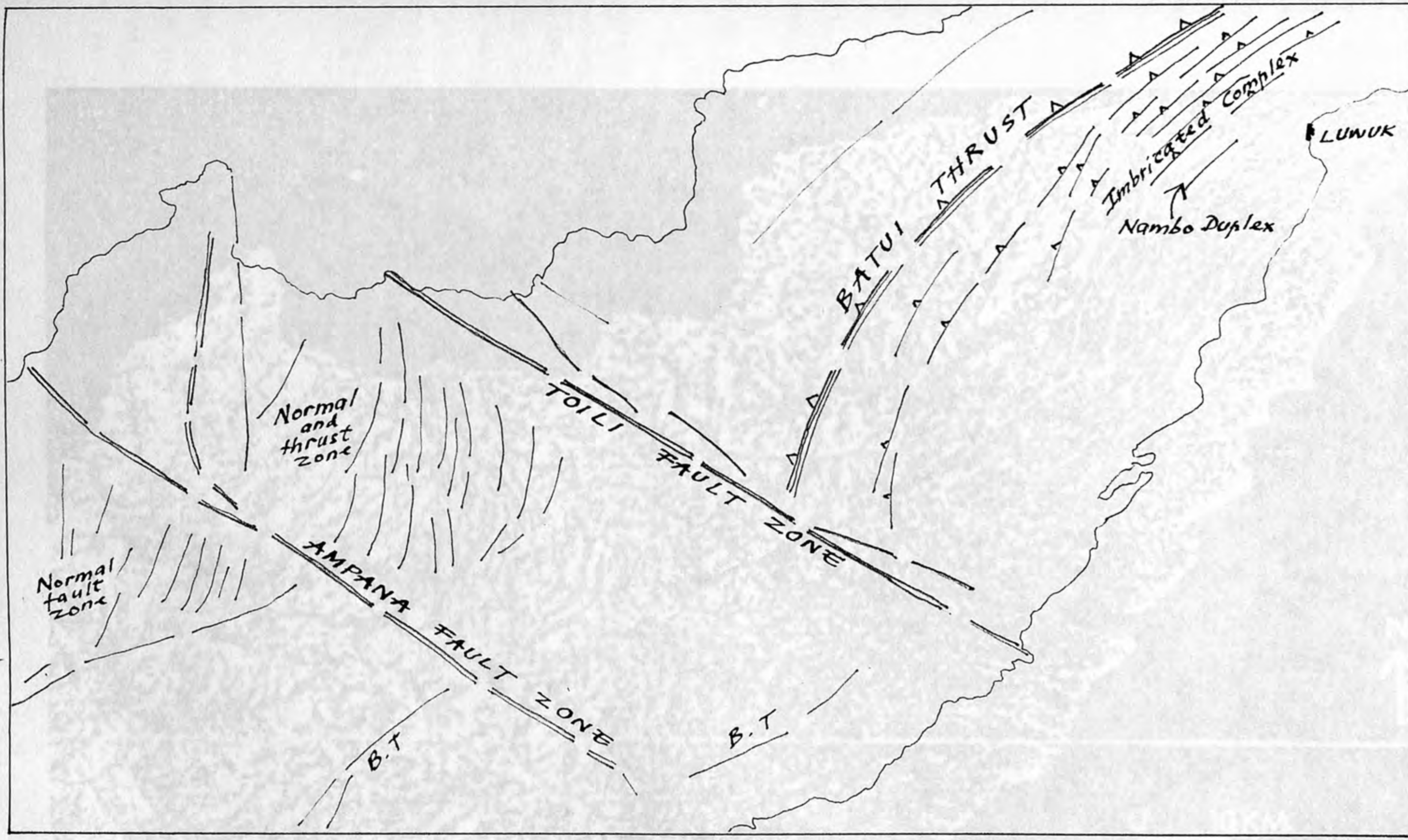
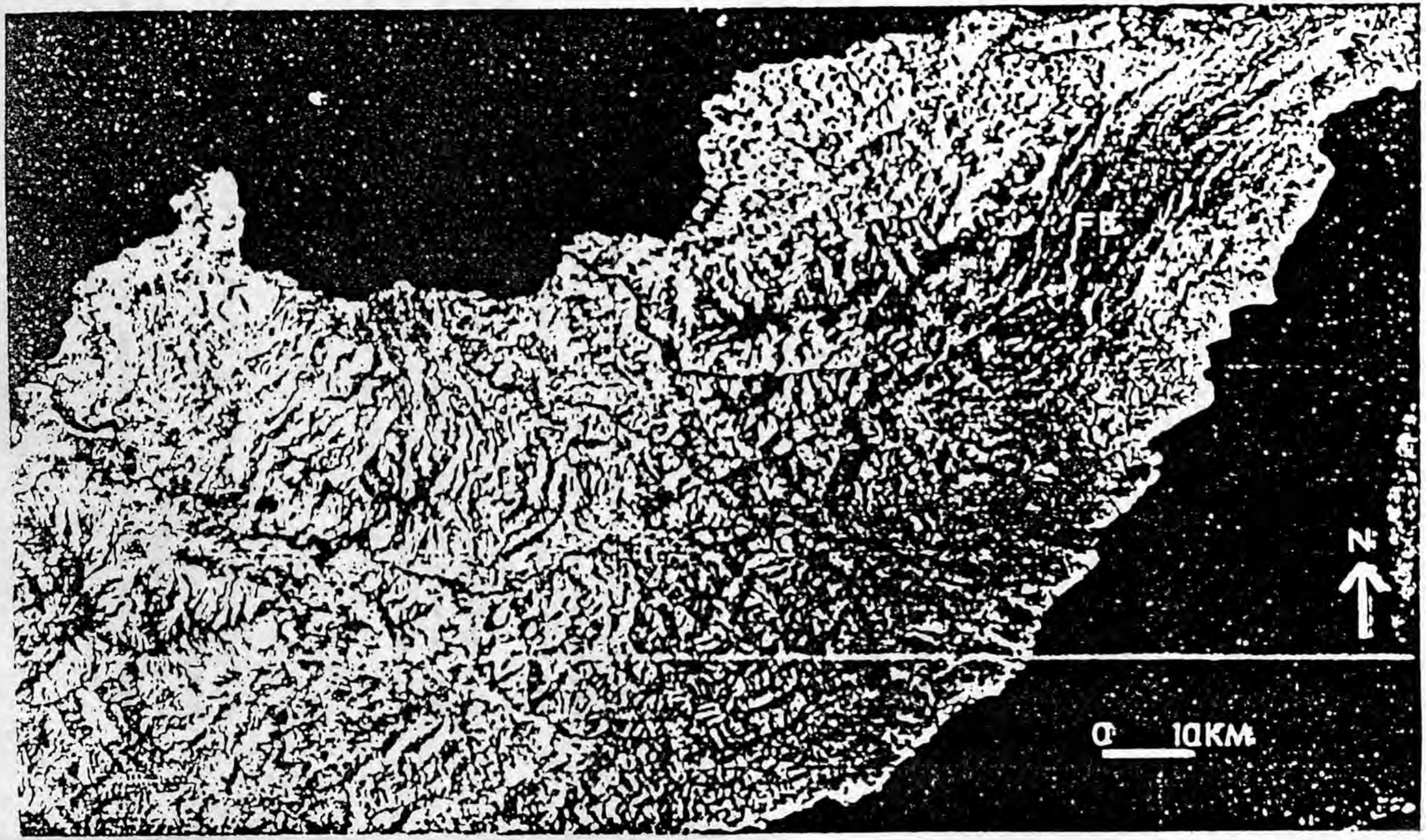


Fig. **4.14** Landsat imagery of the East Arm of Sulawesi, showing the northwest trending fault valley along the Toili Fault Zone (TFZ) and Ampana Fault Zone (AFZ).


FB
N
0
10KM

strike-slip fault following the valley of Toili river. The displacement may be at least 20 km. This estimation is based on the distribution of the Salodik Limestones, which occurred in Kolo Atas area and offshore on the floor of Tolo Bay to the south of Toili village, clearly shown in the subsurface data (i.e. Seismic and bore-hole data) and the southward displacement or offset of the Batui Thrust in the region just to the west of the Toili river (Fig. 3.13; 3.14).

Rusmana et al. (1984) and Surono et al. (1984) described coarse clastic sediments (i.e. Batui Group) which have been affected by this fault. Hence, the development of the Toili Fault System postdated the deposition of Batui Group, and must have occurred in Plio-Pleistocene time, which is also the same age as the dextral movement of the Balantak Fault System.

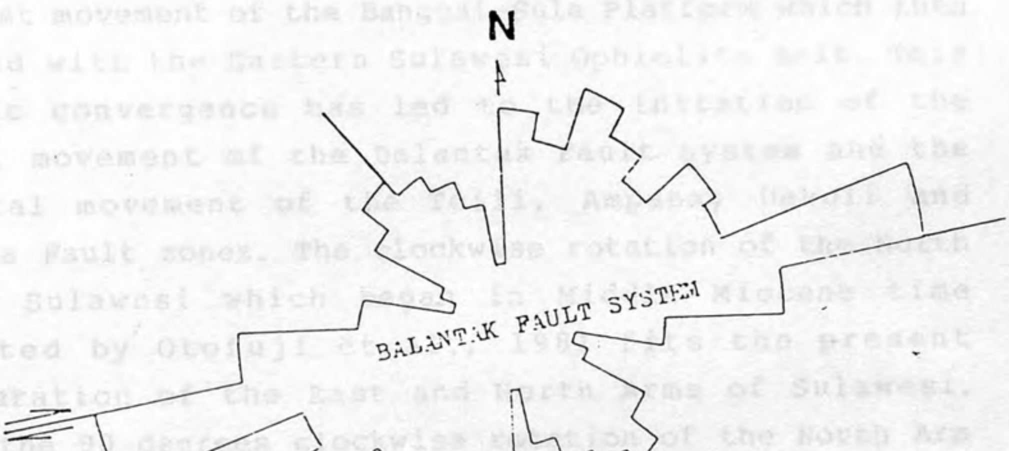
4.4.3 AMPANA FAULT ZONE

The Ampana Fault Zone is interpreted as a dextral strike-slip fault trending in a NNW-SSE direction. The amounts and sense of movement are derived from the separation or displacement of the collision zone in Kolo Atas area and in the Tokala Mountains. The collision zone in these areas is marked by the thrust contact between the continental margin sediments and the ophiolitic rocks. The displacement may be at least 60 km (Fig. 4.16).

4.4.4 UEKULI FAULT ZONE

The Uekuli Fault Zone is a sinistral strike-slip fault, trending in NNW-SSE direction (Fig.4.16). Sukamto and Simandjuntak (1982) interpreted that this fault as continuing to Tolitoli in the upper neck of the North Arm. Hamilton (1979) suggested that this fault zone was a thrust fault.

Fig. 4.15 Rose-diagram showing lineament of steep faults in Poh Head, the East Arm of Sulawesi. The maximum trending in nearly E-W direction may be related to the dextral movement of the Balanatak Fault System. Secondary maxima trends in NW-SE direction due to sinistral movement of strike-slip faults. This pattern is result of pure shear produced by NW-SE compression due to the northwest movement of the Banggai-Sula Platform.



The amounts and sense of movement are also based on the apparent displacement of the thrust-contacts of the continental margin sequence with the ophiolitic rocks. Recent geological maps of GRDC (Simandjuntak et al., 1981; 1983) show that the thrust contact of the carbonate slope deposits of Triassic Tokala Formation with the ultramafic rocks in eastern part of Central Sulawesi is displaced from that in Tokala Mountains for at least 70 km. The sense of movement appears to be left lateral (Fig. 4.16).

Initiation of these two strike-slip faults is essentially related to the northwest-movement of the Banggai-Sula Platform which collided with the ophiolite suit. The structural analysis above, however, indicates that the present structural and tectonic configuration of the East Arm of Sulawesi is produced essentially by the northwest movement of the Banggai-Sula Platform which then collided with the Eastern Sulawesi Ophiolite Belt. This tectonic convergence has led to the initation of the dextral movement of the Balantak Fault System and the sinistral movement of the Toili, Ampana, Uekuli and Minahasa Fault zonez. The clockwise rotation of the North Arm of Sulawesi which began in Middle Miocene time suggested by Otofuji et al., 1981 fits the present configuration of the East and North Arms of Sulawesi. Hence, the 90 degrees clockwise rotation of the North Arm is not necessary or is not the only requirement for producing the present configuration of the East and North Arms of Sulawesi. This aspect will be further discussed in Chapter 5.

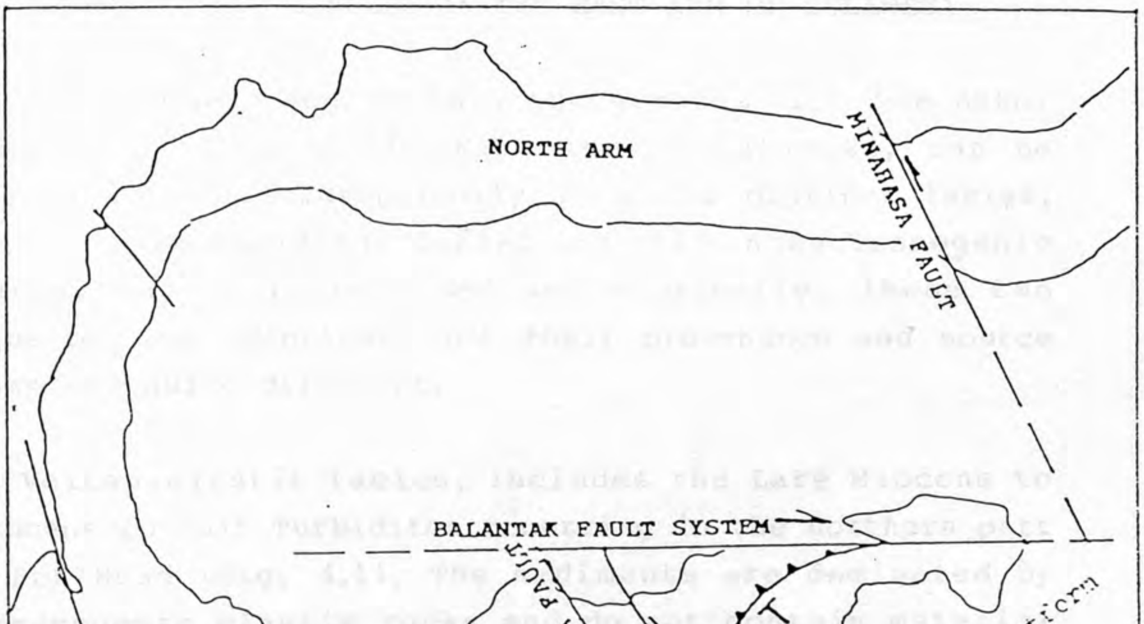


Fig. 4.16 Map showing structural configuration of the East Arm of Sulawesi. The development of dextral movement of the Balantak Fault System essentially results from pure shear, which produces by NW-SE compression due to the northwest movement of the Banggai-Sula Platform.

4.5 REGIONAL SEDIMENTATION PATTERNS OF POST OROGENIC COARSE CLASTIC ROCKS IN THE EAST ARM OF SULAWESI

The Neogene sedimentary successions (i.e. the Batui Group) occurring in the East Arm of Sulawesi, can be divided lithostratigraphically into two distinct facies, i.e. (i) volcaniclastic facies and (ii) nonvolcanogenic coarse clastic facies. Sedimentologically, these two sequences are identical, but their provenance and source areas are quite different.

(i) Volcaniclastic facies, includes the Late Miocene to Pliocene Lonsuit Turbidites occurring in the northern part of Poh Head (Fig. 4.1). The sediments are dominated by volcanogenic clastic rocks and do not contain material derived from the collision complex of the East Arm of Sulawesi. Planktonic foraminifera of Late Miocene to Pliocene age occurs in intercalated beds of calcareous sandstone. Palaeocurrent data suggest that the volcanic debris was derived from the northwest. They are product of sediment gravity flows and deposited in the inner fan system in the Binsil Subbasin, considered to have been part of the Tomini Trough in Late Miocene time.

(ii) Nonvolcanogenic coarse clastic facies includes the Late Miocene to Pliocene Kolo Beds, Biak Conglomerates, Kintom Formation (Surono et al., 1984) and Bongka Formation (Simandjuntak et al., 1983). The first three units occur in the southern part of the East Arm, while the Bongka Formation occupies the northern part of central East Arm (Fig. 4.1). These rocks are typically post-orogenic coarse clastic sediments as indicated by the presence of material which is wholly derived from the collision complex in the East Arm of Sulawesi. The presence of planktonic foraminifera in these rocks indicates that they were deposited in an open marine

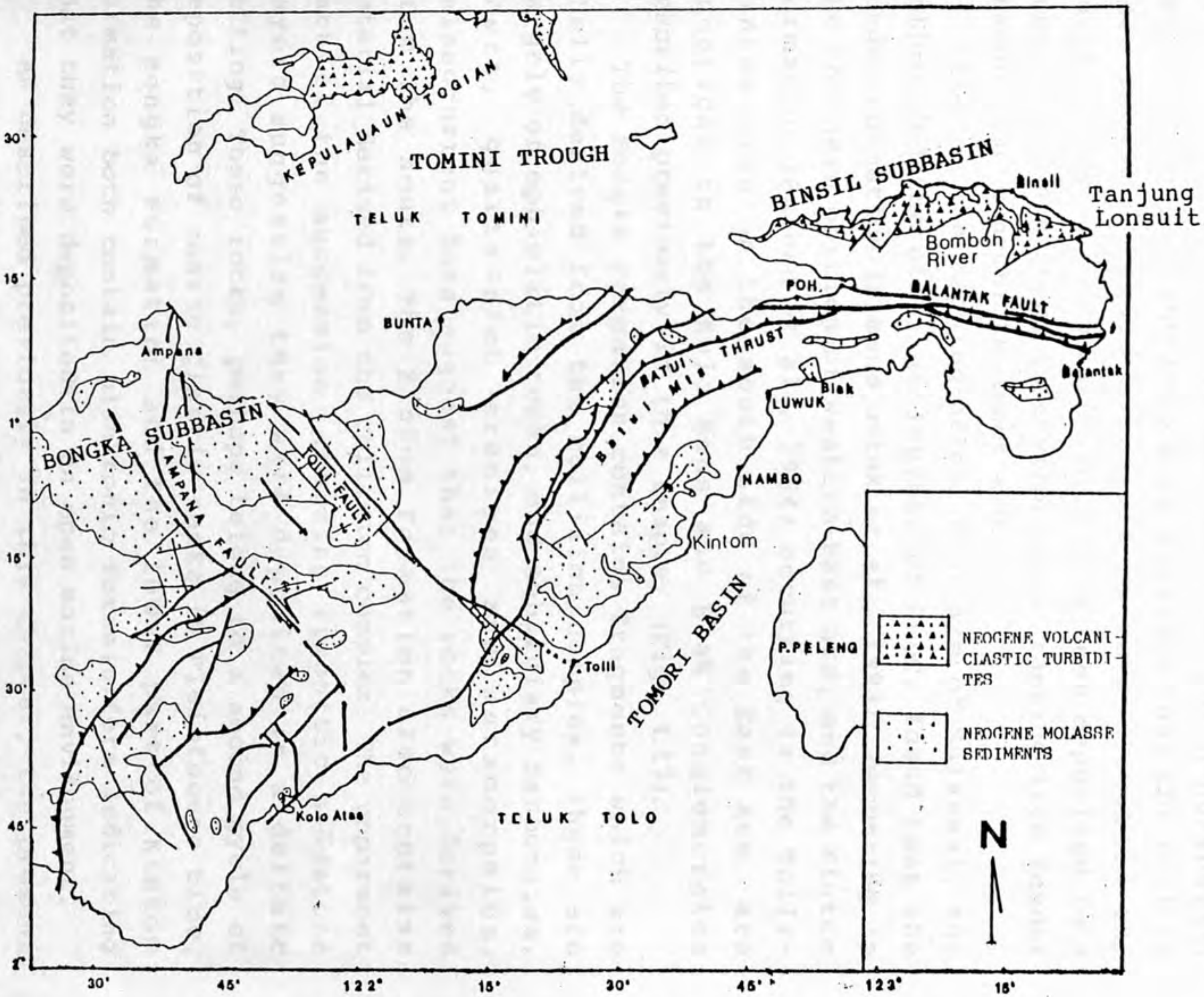


Fig. 4.17 Map showing the distribution of the Neogene coarse clastic and volcanogenic sediments.

depositional setting. Palaeocurrent data suggest that the Biak Conglomerates were derived from north-northwest. Sedimentology of the Kolo Beds suggests that the unit is more proximal toward north and is more distal to the south. They are interpreted to have been deposited in a submarine fan system, in environments ranging from feeder channel (or canyon) to inner fan.

During the mapping of the East Arm of Sulawesi, the author and the other geologists of GRDC, found that the Bongka Formation (Simandjuntak et al., 1983), occurring in the northern portion of western East Arm, and the Kintom Formation (Surono et al., 1984) occurring in the Toili-Kintom area on the south side of the East Arm are identical to the Kolo Beds and Biak Conglomerates described previously in this chapter (Fig. 4.17).

The Bongka Formation contains fragments which are wholly derived from the collision complex, these are largely of ophiolitic rocks, and subsidiary carbonates, chert, quartz-rich arenites and metamorphics. Palaeocurrent data suggest that the rocks were derived from the south. The Kintom Formation also contains material derived from the collision complex. The uppermost part of the succession contains limonitic sandstone layers suggesting they were deposited in a deltaic setting. These rocks, perhaps belong to a second cycle of deposition of coarse clastic rocks in Pleistocene time. The Bongka Formation and the lower part of Kintom Formation both contain planktonic foraminifera indicating that they were deposited in an open marine environment.

As described previously in this chapter, the present physiographic and structural configuration of the East Arm and the distribution of the Neogene coarse clastic sediments appear to have been affected and modified by the continuously active or reactivated tectonic convergence between the Banggai-Sula Platform and Eastern Sulawesi Ophiolite Belt since Middle Miocene time and the

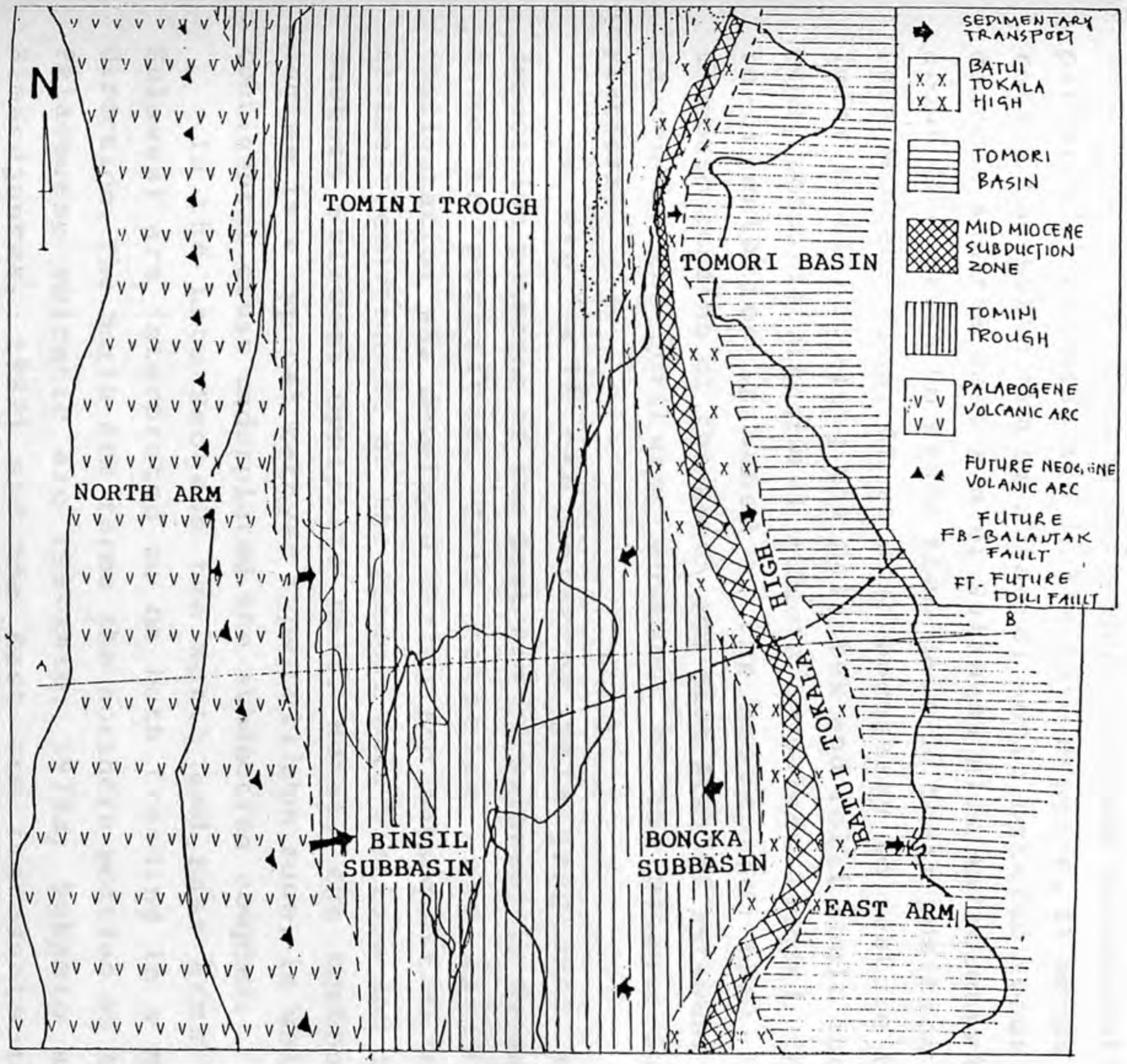


Fig. 4.18 Map showing the inferred palaeogeography and sedimentation patterns of Neogene coarse clastic sediments in the East Arm of Sulawesi.

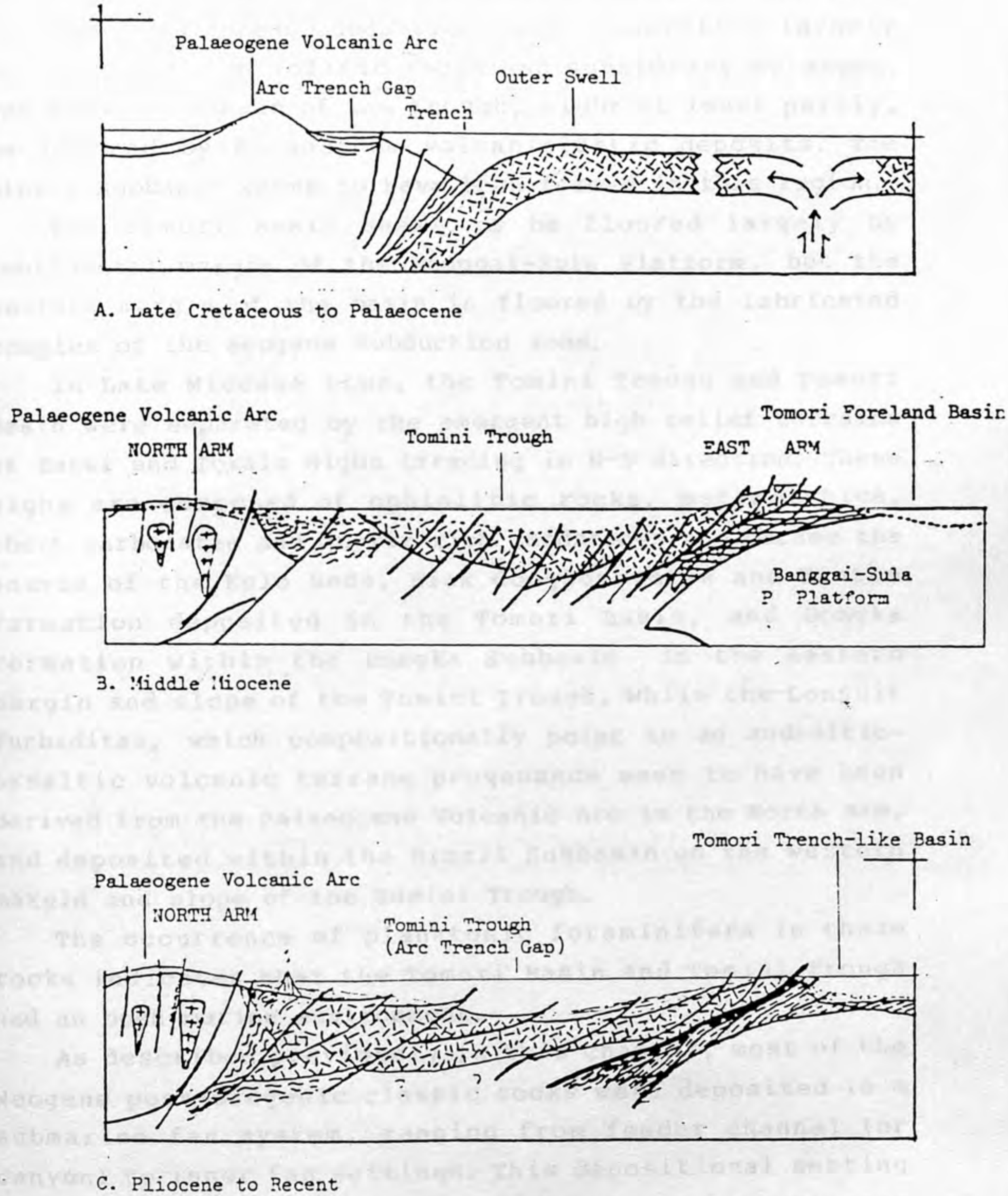
initiation and development of the strike-slip faults in the East Arm of Sulawesi. Tectonic analysis of this region is discussed in Chapter 5.

It is assumed that the inferred physiography of the Palaeogene Volcanic Arc and the Late Cretaceous subduction zone was trending in N-S direction. In attempting to reconstruct the inferred palaeogeography and sedimentation patterns of the East Arm in the Late Neogene, it needs to rotate the North Arm of Sulawesi anticlockwise about 90 degrees and the East Arm of Sulawesi also anticlockwise about 45 degrees to place them back to their position in Late Miocene time. It is also necessary to restore the inferred offset along the Balantak and Toili fault zone. Other possible structural affects have been ignored. This re-orientation is based on the assumption that the tectonic bending of the North and East Arms of Sulawesi is caused by the northwest movement of the Banggai-Sula Platform.

In Fig. 4.18 the inferred paleogeography and deposition patterns of the East Arm of Sulawesi in Neogene time is portrayed. This tectonic and basinal configuration was developed during and subsequent to the tectonic convergence of the Banggai-Sula Platform and the Eastern Sulawesi Ophiolite Belt. Hence, the tectonic regime is a typical Tethyan type collision zone, in which continental crust underplated the subduction complex.

In the Late Neogene the North and East Arms of Sulawesi are interpreted to be both trending in a N-S direction. The North Arm forms the northern portion of the Palaeogene volcanic arc (Sukamto, 1975a; Sukamto and Simandjuntak, 1982) and the East Arm represents the northern portion of the Eastern Sulawesi Ophiolite Belt. The North Arm was separated from the East Arm by the Tomini Basin or Trough, and the Tomori Basin developed to the east of the East Arm. The Tomori Basin originally developed as a foreland basin, elongated in N-S direction,

Fig. 4.19 Map showing the inferred palaeogeography, palaeo-basinal setting and depositional patterns of the East Arm in Neogene times.



while the Tomini Basin developed as an Outer Arc Trough, also with a N-S trend.

The Tomini Trough, which in this stage represent the northern portion of the Neogene outer arc trough, appears to be floored by the accretionary prism of the Late Cretaceous-Paleocene subduction zone, consisting largely of imbricated ophiolitic rocks and subsidiary melanges. The western margin of the trough, might at least partly, be floored by Palaeogene volcaniclastic deposits. The Binsil Subbasin seems to have been formed in this region.

The Tomori Basin seems to be floored largely by continental margin of the Banggai-Sula Platform, but the western margin of the basin is floored by the imbricated complex of the Neogene subduction zone.

In Late Miocene time, the Tomini Trough and Tomori Basin were separated by the emergent high relief terrains of Batui and Tokala Highs trending in N-S direction. These highs are composed of ophiolitic rocks, metamorphics, chert carbonates and quartz-rich sandstones and became the source of the Kolo Beds, Biak Conglomerates and Kintom Formation deposited in the Tomori Basin, and Bongka Formation within the Bongka Subbasin in the eastern margin and slope of the Tomini Trough. While the Lonşuit Turbidites, which compositionally point to an andesitic-basaltic volcanic terrane provenance seem to have been derived from the Palaeogene Volcanic Arc in the North Arm, and deposited within the Binsil Subbasin on the western margin and slope of the Tomini Trough.

The occurrence of planktonic foraminifera in these rocks indicates that the Tomori Basin and Tomini Trough had an open marine environment.

As described previously in this chapter, most of the Neogene post-orogenic clastic rocks were deposited in a submarine fan system, ranging from feeder channel (or canyon) to inner fan settings. This depositional setting was accommodated by the evolution of the Late Miocene

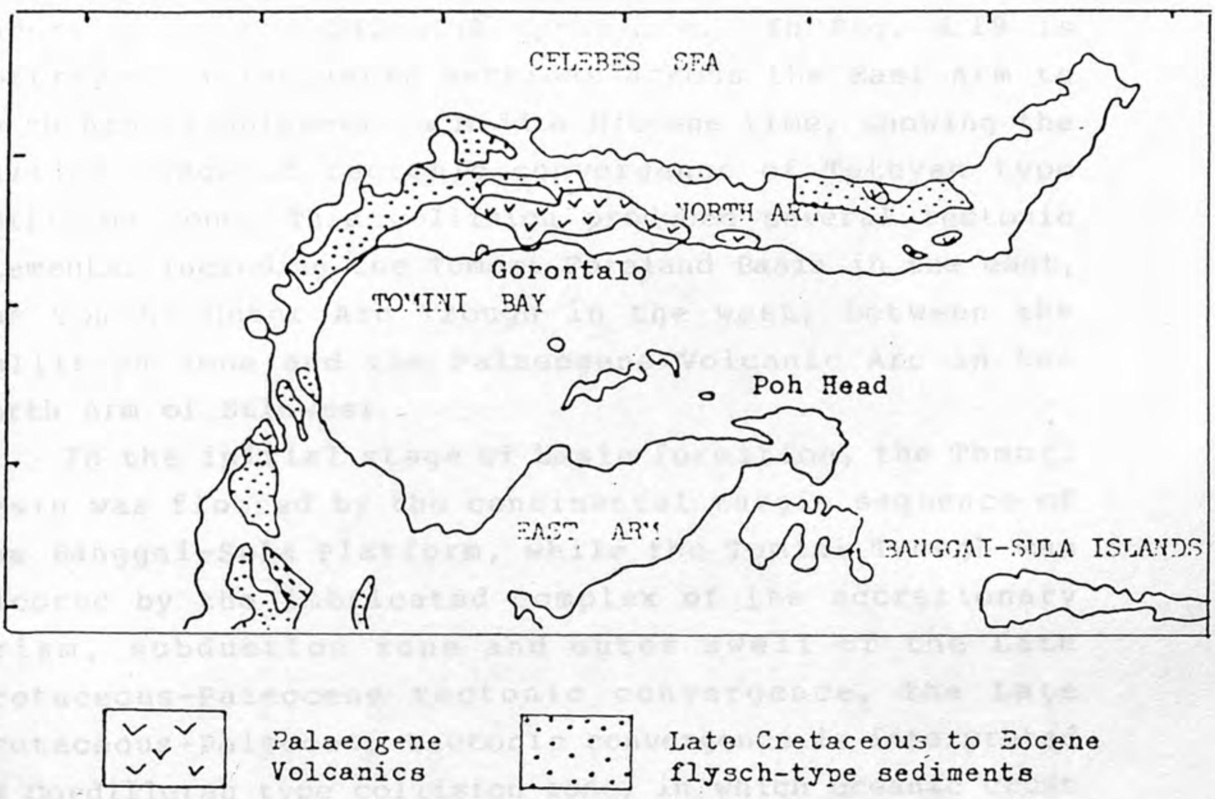


Fig. 4.19.1 Map showing the occurrence of Palaeogene Volcanics in the North Arm of Sulawesi, which are interpreted to be source of the Lonsuit Turbidites.

Basinal setting, from foreland basin to trench-like basin of the Tomori Basin, and from outer arc trough to arc-trench gap basin of the Tomini Trough.

The evolution of these basins is essentially due to the continuously active collision zone in the East Arm of Sulawesi, in which the Banggai-Sula Platform is underplating the Balantak Ophiolite. In Fig. 4.19 is portrayed palinspastic sections across the East Arm to North Arm of Sulawesi in Middle Miocene time, showing the initial stage of tectonic convergence of Tethyan type collision zone. This collision produced several tectonic elements, including the Tomori Foreland Basin in the east, the Tomini Outer Arc Trough in the west, between the collision zone and the Palaeogene Volcanic Arc in the North Arm of Sulawesi.

In the initial stage of basin formation, the Tomori Basin was floored by the continental margin sequence of the Banggai-Sula Platform, while the Tomini Trough was floored by the imbricated complex of the accretionary prism, subduction zone and outer swell of the Late Cretaceous-Paleocene tectonic convergence. The Late Cretaceous-Paleocene tectonic convergence is interpreted as Cordilleran type collision zone, in which oceanic crust was subducted westwards beneath the island arcs of the Sundaland of SE Asia. This assumption is based on the occurrence of Palaeogene volcanic arc and Late Cretaceous flysch-type sediments in the Western Sulawesi Volcano-Plutonic Belt (WSVB) juxtaposed or in fault contact with the Central Sulawesi Metamorphic Belt (CSMB) and Eastern Sulawesi Ophiolite Belt (ESOB) to the east (Simandjuntak, 1980; Sukamto and Simandjuntak, 1982).

In Late Miocene to Pliocene time, as the collision was continuously active or reactivated, the tectonic configuration of the region was modified and gave rise to the development of Tomori Basin to become a trench-like basin, and the Tomini Trough became an arc-trench gap

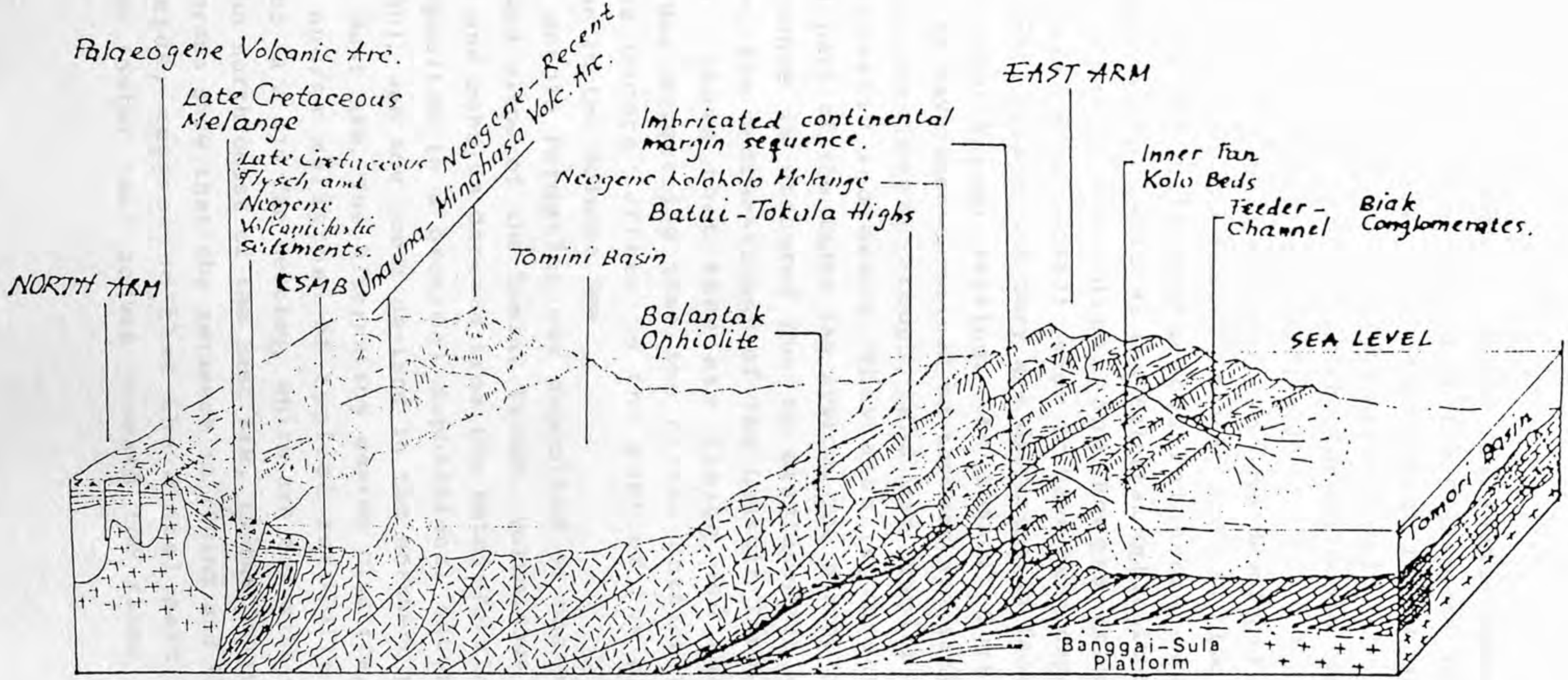


Fig. 4.20 Block Diagram showing tectonic convergence and basinal setting in the East Arm of Sulawesi in Neogene time.

basin. A Neogene volcanic arc (i.e. Minahasa-Unauna Arc) developed and was superimposed on the Palaeogene Volcanic Arc in the western margin of the Tomini Trough and the uplifting of Neogene accretionary prism forming the Batui-Tokala Highs, which separated the Tomori Basin from the Tomini Trough (Fig. 4.18; 4.19C; 4.20).

The Binsil Subbasin appears to have developed on the western margin of the Tomini Trough, and was floored by outer swell of Late Cretaceous-Paleocene tectonic convergence, which now forms the Balantak Ophiolite.

The Lonsuit Turbidites and Bongka Formation (Simandjuntak et al., 1983) seem to have been deposited in the Tomori Trough, but they are different in facies and site of depositional setting. The Lonsuit Turbidites appears to have been deposited on the western margin and slope of the Tomori Trough, and are dominated by volcaniclastic sediments. They also represents the proximal part of the inner fan system. The distal part of the sequence is obscured due to displacement of the sequence. The sedimentology of the Lonsuit Turbidites, however, shows that they are fining and thinning eastwards, suggesting that the distal portion of the sequence occurs further to the east of the present shoreline in the Molucca Sea.

The Bongka Formation was deposited on the eastern margin and slope of the Tomini Trough. Voluminous coarse pebbles and cobbles derived from the Batui-Tokala Highs were deposited in a proximal depositional setting (or inner fan), and now occur on-land in the northern part of western East Arm. The alternating coarse clastics, silty shales and/or marlstone of typical turbidites were deposited in a mid fan setting, which are now well exposed along the north coast of the East Arm, in the Bongka area. The outcrops show that the sequence is fining and thinning northwards, suggesting that the distal part of the sequence (outer fan) occurs beneath the floor of the

Tomini Bay, further to the north.

In the fan system, the sediments were introduced into the basin from submarine canyons onto a large complex of fans that extended out into basin. In the East Arm of Sulawesi, the Biak Conglomerates are considered sedimentologically to have been deposited in a submarine canyon-feeder channel (canyon) system and the Kolo Beds in an upper fan system.

A similar depositional setting is well-described and documented from elsewhere, such as Hamilton (1967), for deposits in Gulf of Alaska, Normark (1970) for development of the San Lucas deep sea fan, Hanner (1971) for deposition and development of the Redondo Submarine fan, southern California, Henderson (1972) for deposits in Yellowknife , Northwest Territories, Canada, Hampton (1973) for development of the La Jolla Submarine fan, California and Walker and Mutti (1973) for a facies model of submarine fan deposits. In all cases, the submarine canyons (or feeder canyons) cut the continental margins or other terranes and have diverse origins including direct river, turbidity current erosion, slumping and erosion of faulted gullies (Leeder, 1983).

CHAPTER 5

PALAEOGEOGRAPHIC RECONSTRUCTION AND TECTONIC ANALYSES OF THE EAST ARM OF SULAWESI

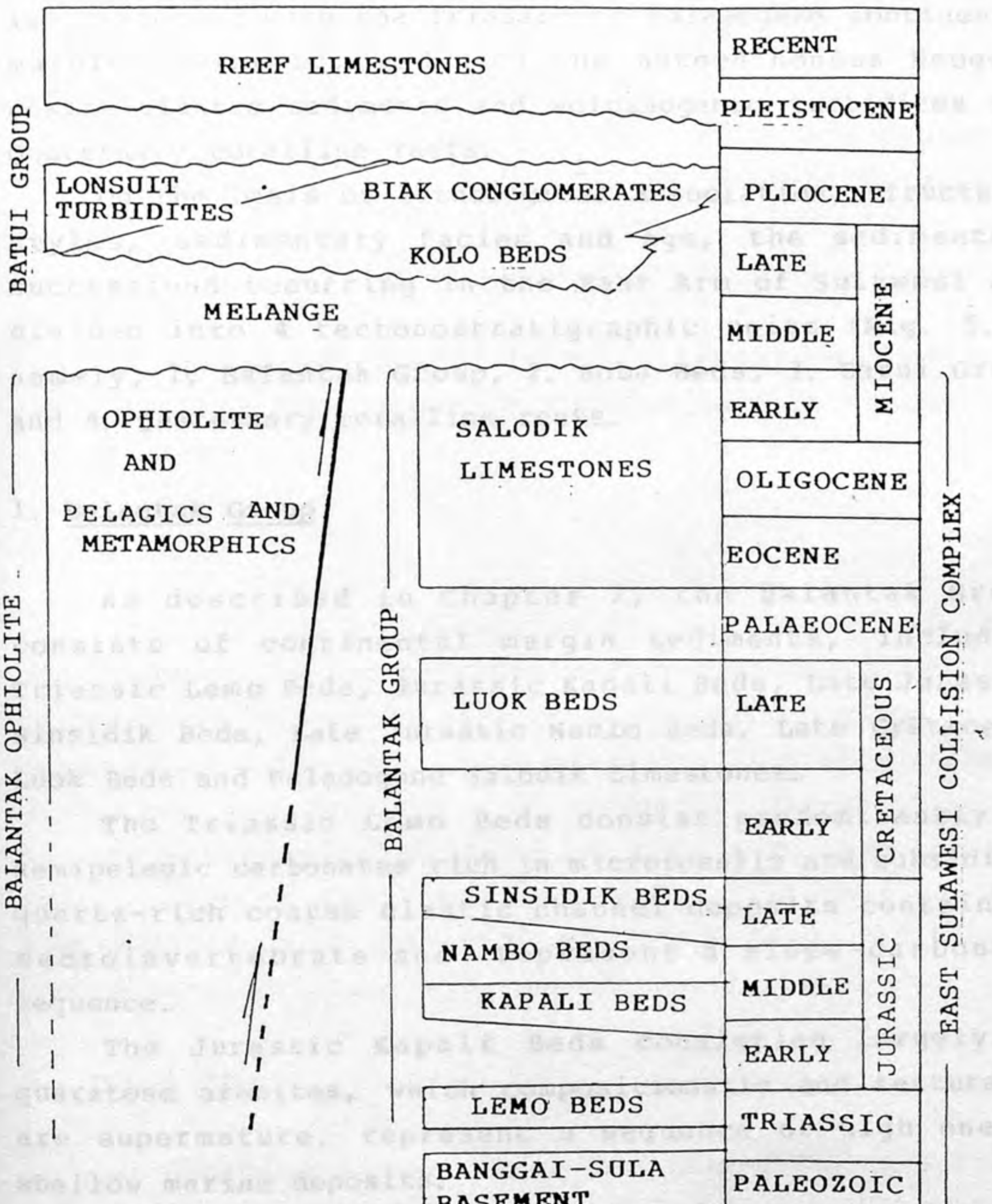
5.1 INTRODUCTION

This Chapter deals with an analysis of the palaeogeography and tectonic evolution of the East Arm of Sulawesi. The analysis is based on data presented in Chapters 1, 2, 3, and 4. Data collected during the mapping of the region by the author and his colleagues from GRDC are included as well as information obtained from previous work.

The geology of the east Arm of Sulawesi is examined in order to reconstruct the tectonostratigraphy of the region. By understanding the tectonostratigraphy of the collision complex in the East Arm of Sulawesi, the palaeogeographic reconstruction and the tectonic evolution of the region can be analysed more accurately.

Several alternative kinematic models for the evolution of the eastern Sulawesi collision zone are discussed, in order to understand the regional context and kinematics of the eastern Sulawesi collision zone. The following section will describe and discuss the geology and structures of the East Arm of Sulawesi, and review the geology of the Banggai-Sula, Western Sulawesi Volcano-Plutonic Belt (WSVB) and the Central Sulawesi Metamorphic Belt (CSMB). The focus will be, particularly on the nature of structural elements, the age, facies and inferred depositional setting of the accreted sedimentary successions and the age of the collision episodes.

Fig. 5.1 TECTONOSTRATIGRAPHY OF THE EAST ARM OF SULAWESI.



5.2 TECTONOSTRATIGRAPHY OF THE EAST ARM OF SULAWESI

As described previously in Chapter 4, the collision complex in the East Arm of Sulawesi consists of two distinctive structural domains, i.e. (i) the imbricated complex containing an allochthonous ophiolite belt, which is juxtaposed with the Triassic to Palaeogene continental margin sediments, and (ii) the autochthonous Neogene coarse clastic sediments and volcanogenic turbidites and Quaternary coralline reefs.

On the basis of lithological association, structural styles, sedimentary facies and age, the sedimentary successions occurring in the East Arm of Sulawesi are divided into 4 tectonostratigraphic units (Fig. 5.1), namely, 1. Balantak Group, 2. Boba Beds, 3. Batui Group and 4. Quaternary coralline reefs.

1. Balantak Group

As described in Chapter 2, the Balantak Group consists of continental margin sediments, including Triassic Lemo Beds, Jurassic Kapali Beds, Late Jurassic Sinsidik Beds, Late Jurassic Nambo Beds, Late Cretaceous Luok Beds and Palaeogene Salodik Limestones.

The Triassic **Lemo Beds** consist predominantly of hemipelagic carbonates rich in microfossils and subsidiary quartz-rich coarse clastic channel deposits containing macroinvertebrate and represent a slope-carbonate sequence.

The Jurassic **Kapali Beds** consisting largely of quarztose arenites, which compositionally and texturally are supermature, represent a sequence of high energy shallow marine deposits.

The Late Jurassic **Sinsidik** Beds, which consist of alternating glauconitic calcarenite and argillaceous limestones with microfossils and macroinvertebrate, were deposited in neritic depth.

The Late Jurassic **Nambo Beds**, consisting predominantly of marlstone and subsidiary carbonaceous shales with macroinvertebrates and microfossils, represent a carbonate sequence deposited in outer neritic depth.

The Late Cretaceous **Luok Beds**, which are dominated by calcilutite and/or lithified calcareous ooze with chert nodules and rich in microfossils, represent an upper slope-carbonate sequence.

The Palaeogene **Salodik Limestones** are typically carbonate platform deposits rich in foraminifera, both benthic and the planktonic. Texturally and compositionally, the unit was deposited on a deepening shelf.

All these units are highly tectonised and occur in fault-slivers forming the imbricated complex in the East Arm of Sulawesi.

2. Boba Beds

As described previously in Chapter 3, the Boba Beds, which are dominated by radiolarian chert and subsidiary calcilutites represent pelagic sediments deposited on the ocean floor below CCD, which possibly fluctuated in depth with time. The occurrence of syngenetic manganese oxides and the well-bedded nature of the Boba Beds, make the unit quite distinct from the Luok Beds. This unit is also highly deformed, highly faulted and thrusted, and usually occurs in association with the ophiolite rocks.

3. Batui Group

In Chapter 4, it is shown that this group consists of two distinctive sedimentary sequences, i.e. (i) Neogene post-orogenic coarse clastic rocks, including the Kolo Beds and the Biak Conglomerates, which consist essentially of sediments derived from the collision complex, and (ii) Neogene Lonsuit Turbidites, which are typically volcaniclastic sediments which do not contain material derived from the collision complex.

Both sequences were deposited in a submarine fan depositional setting. This group is only gently folded, and faulted in many places.

4. Quaternary Coralline Reefs

This assemblage consists of an only slightly deformed carbonate sequence dominated by Quaternary coralline reefs. Reefs occur in at least three terraces at approximately 400 m, 75 m and 20 m above sea level along the south coast of the East Arm. The amount of uplift declines gradually westwards until the reefs disappear below sea-level.

5.3 GEOLOGICAL FRAMEWORK OF THE EASTERN SULAWESI COLLISION COMPLEX

The collision zones in the East Arm of Sulawesi, and/or to a large extent in the whole of eastern Sulawesi, are characterised by the accretion of a distinctive sequence of Mesozoic to Miocene sediments. These rocks are largely carbonates deposited in deep continental margin to platformal settings. Identical palaeogeographic settings of these accreted rocks can be inferred on the basis of comparison and correlation with the Banggai-Sula and other platforms (or microcontinents), in eastern Indonesia. In the following section, the geology of Banggai-Sula Platform (BSP), the Western Sulawesi Volcano-Plutonic Belt (WSVB) and the Central Sulawesi Metamorphic Belt (CSMB) is reviewed.

A. BANGGGAI-SULA PLATFORM

The Banggai-Sula Platform forms an E-W chain of islands located on a shallow triangular shaped platform that extends for some 500 km to the east of Sulawesi, and separates the Maluku Sea to the North from the Banda Sea to the southwest, and Ceram Sea to the east. This platform, has previously been called the Sula Spur (Stille, 1945), Banggai-Sula mini (micro) continent (Hamilton, 1979) and the Sula Platform (Silver, 1981; Smith, 1983). In this study, the Banggai and Sula Islands are considered to form a single platform.

The Banggai-Sula Platform is composed of a pre-Jurassic metamorphic and plutonic basement complex overlain by Jurassic through Tertiary shelf sediments (Van Bemmelen, 1949; Sukamto, 1975a; Hamilton, 1979; Surono and Sukarna, 1985; Supanjono and Haryono, 1985). The basement complex consists largely of metamorphic rocks, including slate, phyllite, and quartzite with subsidiary epidote-chlorite schists, garnet-kyanite schist, amphibolite, gneiss and hornfels. These metamorphics are intruded by Permo-Triassic (217 - 245 my.) granitoids (Sukamto, 1975a; Surono and Sukarna, 1985; Supanjono and Haryono, 1985). Rhyolitic volcanics (Mangole Volcanics, Surono & Sukarna, op.cit.) overlie the metamorphic and plutonic rocks. Two Rb-Sr whole-rocks analyses on rhyolite give ages of 210 $\pm$ 25 my. and 330 $\pm$ 90 my. (Sukamto, op.cit.), which are consistent with the stratigraphic constraint. Hamilton (1979) considered that the volcanic rocks were co-magmatic with the granitoid rocks (Fig. 5.2).

The basement complex is unconformably overlain by polymictic and arkosic conglomerates, arkose and quartzite of the early Jurassic Kabauw and Bobong Formations, which are conformably overlain by the Middle to Late Jurassic Buya Formation, consisting of black calcareous and non-calcareous shales and mudstones (Fig. 5.3). The Buya

Formation is paraconformably overlain by Late Cretaceous calcilutite and chalk (Tanamu Formation) on Taliabu and Mangole Islands, and by Eocene carbonate platform deposits (Salodik Formation) on Peleng and Banggai islands (Pigram et al., 1984; Surono & Sukarna, 1985; Supanjono & Haryono, 1985). Late Miocene limestone platform deposits with interbedded quartzose sandstones occur in Banggai and Peleng islands. Quaternary coralline limestone is common throughout the platform.

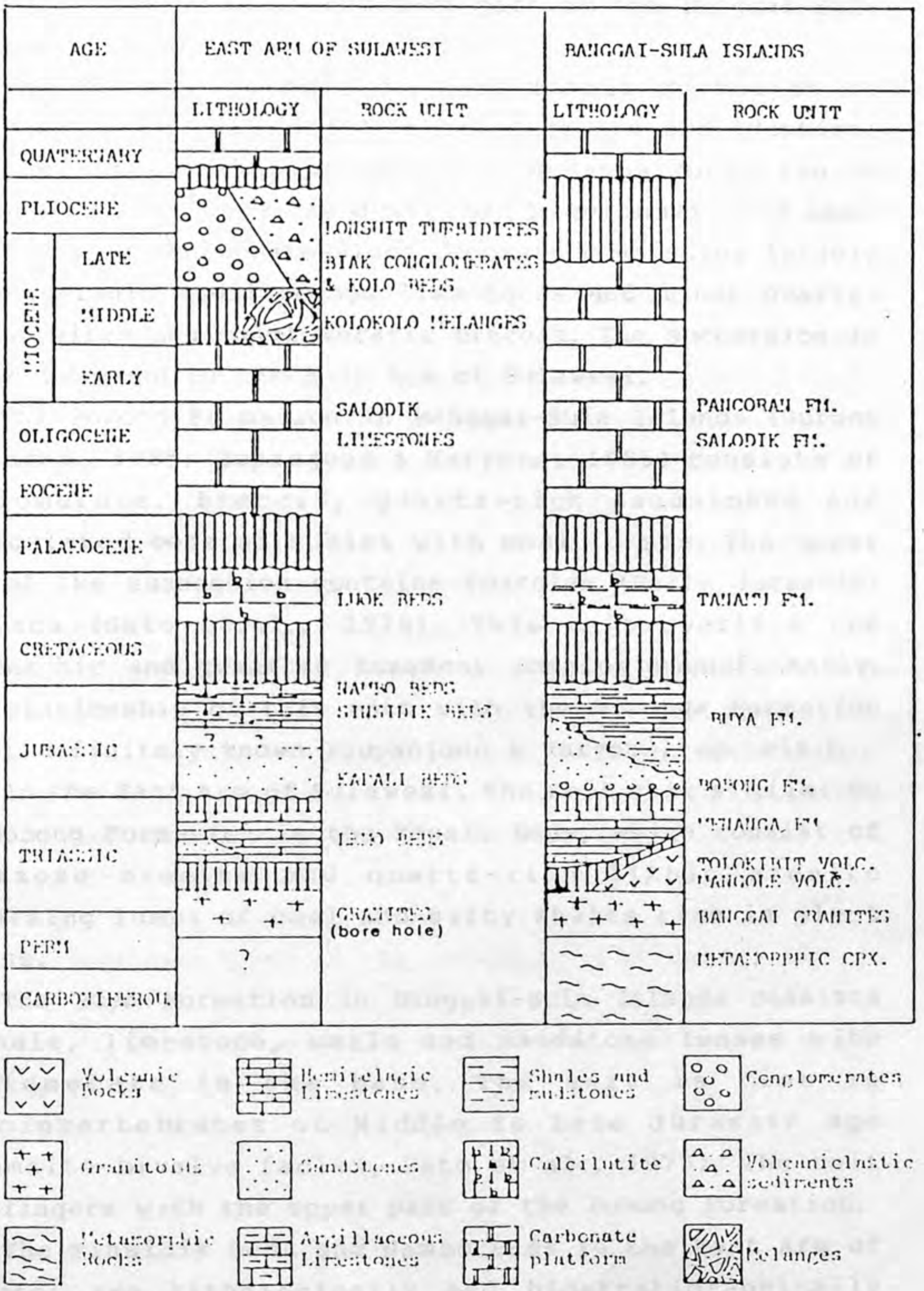
B. Correlation of the Balantak Group with continental margin sequence in Banggai-Sula Islands.

The sedimentology and petrology of the Balantak Group described previously in Chapter 2 shows that most of the rock units are identical to continental margin sediments in Banggai-Sula Islands. The correlation of the Balantak Group with continental margin sediments in Banggai-Sula Islands, which have been mapped recently by the Geological Research and Development Centre (Surono & Sukarna, 1986; Supanjono & Haryono, 1985) is shown in Fig. 5.3.

The Permo-Carboniferous metamorphic rocks are not found on-shore in the East Arm of Sulawesi, but off-shore of Mentawa, borehole data shows that the granitic basement complex occurs at about 800 m depth.

Supanjono & Haryono (op. cit.) described the occurrence of a sequence dominated by limestones and subsidiary lithic arenite, some of which is rich in quartz detritus, and intercalated beds of argillites (i.e. Menanga Formation) in Taliabu Island (the largest island in the Banggai-Sula archipelago). The limestones are thinly bedded, but the arenite beds may up to 2 metres thick. No fossils have been collected from these rocks. The unit is faulted against the Permo-Carboniferous metamorphic rocks; and the outcrops suggest that the unit

Fig. 5.2 Stratigraphic column of the East Arm of Sulawesi and Banggai-Sula Islands (Surono & Sukarna, 1986; Supanjono & Haryono, 1986).



interfingers with the Triassic Tolokibit Volcanics. This unit is not found in the eastern part of the Banggai-Sula Platform (Surono & Sukarna, 1985).

The Triassic Lemo Beds (or Tokala Formation of Simandjuntak, et al., 1983) in the East Arm are identical in lithological association to the Menanga Formation in Banggai-Sula Islands. As described previously, the Lemo Beds represent carbonate-slope deposits consisting largely of hemipelagic argillaceous limestones and minor quartz-rich arenites and conglomeratic breccia. The succession is highly deformed in the East Arm of Sulawesi.

The Bobong Formation in Banggai-Sula Islands (Surono & Sukarna, 1985; Supanjono & Haryono, 1985) consists of conglomerate, breccia, quartz-rich sandstones and intercalated beds of shales with coal lenses. The upper part of the succession contains Toarcian (Early Jurassic) molluscs (Sato et.al., 1978). This unit overlies the metamorphic and granitic basement complex unconformably. The relationship of this unit with the Menanga Formation is not definitely known (Supanjono & Haryono, op. cit.).

In the East Arm of Sulawesi, the rock unit similar to the Bobong Formation is the Kapali Beds, which consist of quartzose arenite and quartz-rich lithic arenite containing lumps of coal and silty shales rich in plant remains.

The Buya Formation in Banggai-Sula Islands consists of shale, limestone, marls and sandstone lenses with conglomerate in the base. The unit is rich in macroinvertebrates of Middle to Late Jurassic age (belemnite-bivalve facies, Sato et al., 1978). The unit interfingers with the upper part of the Bobong Formation.

The Sinsidik Beds and Nambo Beds in the East Arm of Sulawesi are lithologically and biostratigraphically identical to the Buya Formation.

The Tanamu Formation in Banggai-Sula Islands (Surono & Sukarna, 1985) consists of calcilutite or lithified calcareous ooze with intercalated beds of marlstone and calcareous mudstone and attains a thickness of up to 300 metres. The calcilutites are rich in microfossils of Late Cretaceous age. The unit overlies the Buya Formation paraconformably.

In the East Arm of Sulawesi, the Louk Beds are identical to the Tanamu Formation in both lithology and biostratigraphy.

Supanjono & Haryono described the occurrence of Eocene to Middle Miocene carbonates (i.e. Salodik Formation) in the western part of the Banggai-Sula Islands. In the eastern part of the islands, however, only Early to Middle Miocene carbonates with quartz sandstone and conglomerate at the base (i.e. Pancoran Formation) are present (Surono & Sukarna, 1985). No Palaeocene sediments occur in the Banggai-Sula Islands.

In the East Arm of Sulawesi, the Salodik Limestones also contain quartz detritus in the marlstone, but the basal conglomerate and quartz sandstone are not seen.

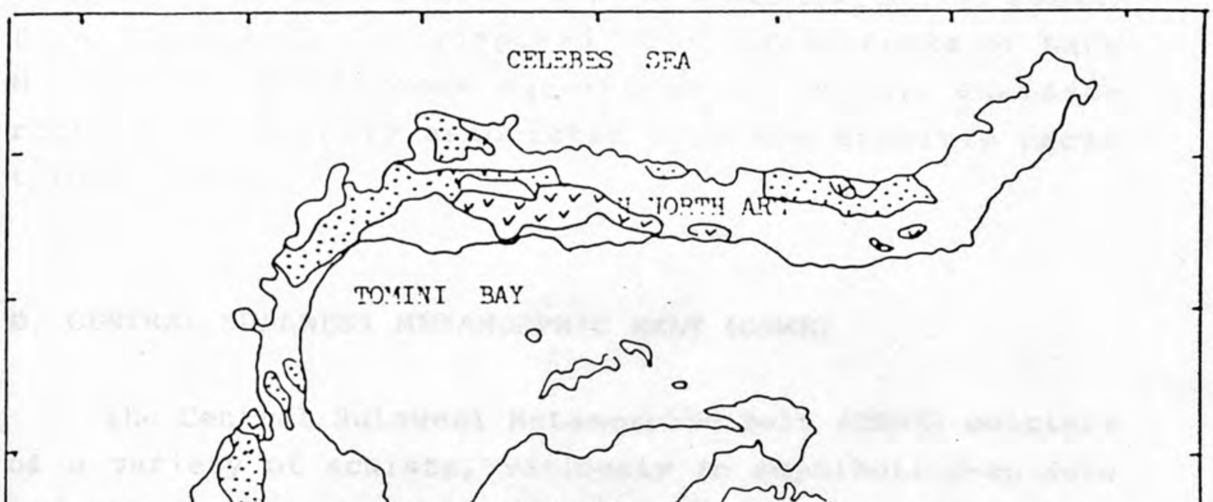
Palaeogene carbonate platform Salodik Limestones accumulated from Eocene to Middle Miocene in the East Arm of Sulawesi and the western part of the Banggai-Sula Islands, but occur much later (i.e. Early Miocene in the eastern part of the islands.

The sedimentology of the Mesozoic continental margin sequence occurring in the Banggai-Sula Islands and the East Arm of Sulawesi suggests that there is a tendency for a deeper or more distal depositional environment to the west, i.e. the East Arm of Sulawesi.

C. WESTERN SULAWESI VOLCANO PLUTONIC BELT (WSVB)

The Western Sulawesi Volcano Plutonic Belt is characterised by Palaeogene volcanics associated with Late Cretaceous to Eocene flysch type sediments, including the

Fig. 5.3 Map showing distribution of Late Cretaceous-Eocene flysch type sediments and Palaeogene Volcanic Arc in the WSVB.



Latimojong Formation and Tinombo Formation of GRDC (Fig. 5.5.1). In many places, Palaeogene Volcanics are intruded by Neogene magmatic rocks and overlain by Neogene Arc Volcanic (Sukamto, 1975a; Sukamto & Simandjuntak, 1982). The magmatic arc is composed of granitic rocks of Late Miocene to Pleistocene age (Sukamto, 1975b). Gneissic rocks occur locally associated with the granitic rocks (Fig. 5.5.2).

D. CENTRAL SULAWESI METAMORPHIC BELT (CSMB)

The Central Sulawesi Metamorphic Belt (CSMB) consists of a variety of schists, variously in amphibolite-epidote and glaucophane-lawsonite (blueschists) or greenschist facies (De Roever, 1947). The belt occupies most of Central Sulawesi, and stretches to the western part of the Southeast Arm and Kabaena Island. Recent geological mapping has shown that the belt also contains phyllite, slate, metachert, marble and serpentinite (Simandjuntak et al., 1981). The age of the metamorphic rocks is not definitely known. Hamilton (1979) suggests a Tertiary age, and Sukamto & Simandjutak (1982) consider that the occurrence of the metamorphic rocks is related to the Late Cretaceous west-dipping subduction zone, in which oceanic crust underplated the eastern continental margin or an island arc offshore the Eurasian continent (Fig. 5.5.2).

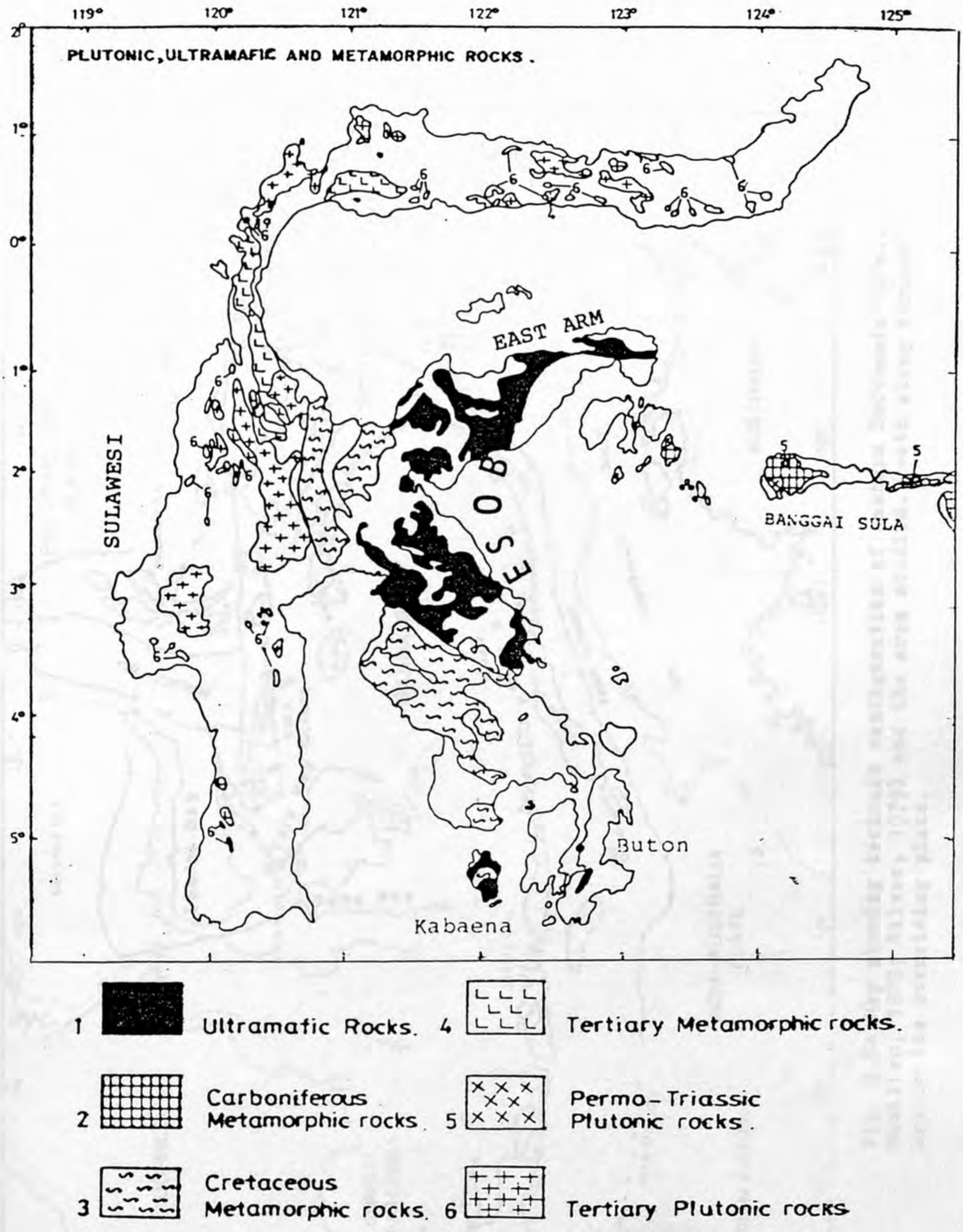


Fig. 5.4 Map showing distribution of plutonic and metamorphic rocks in Sulawesi and Banggai- Sula Islands.

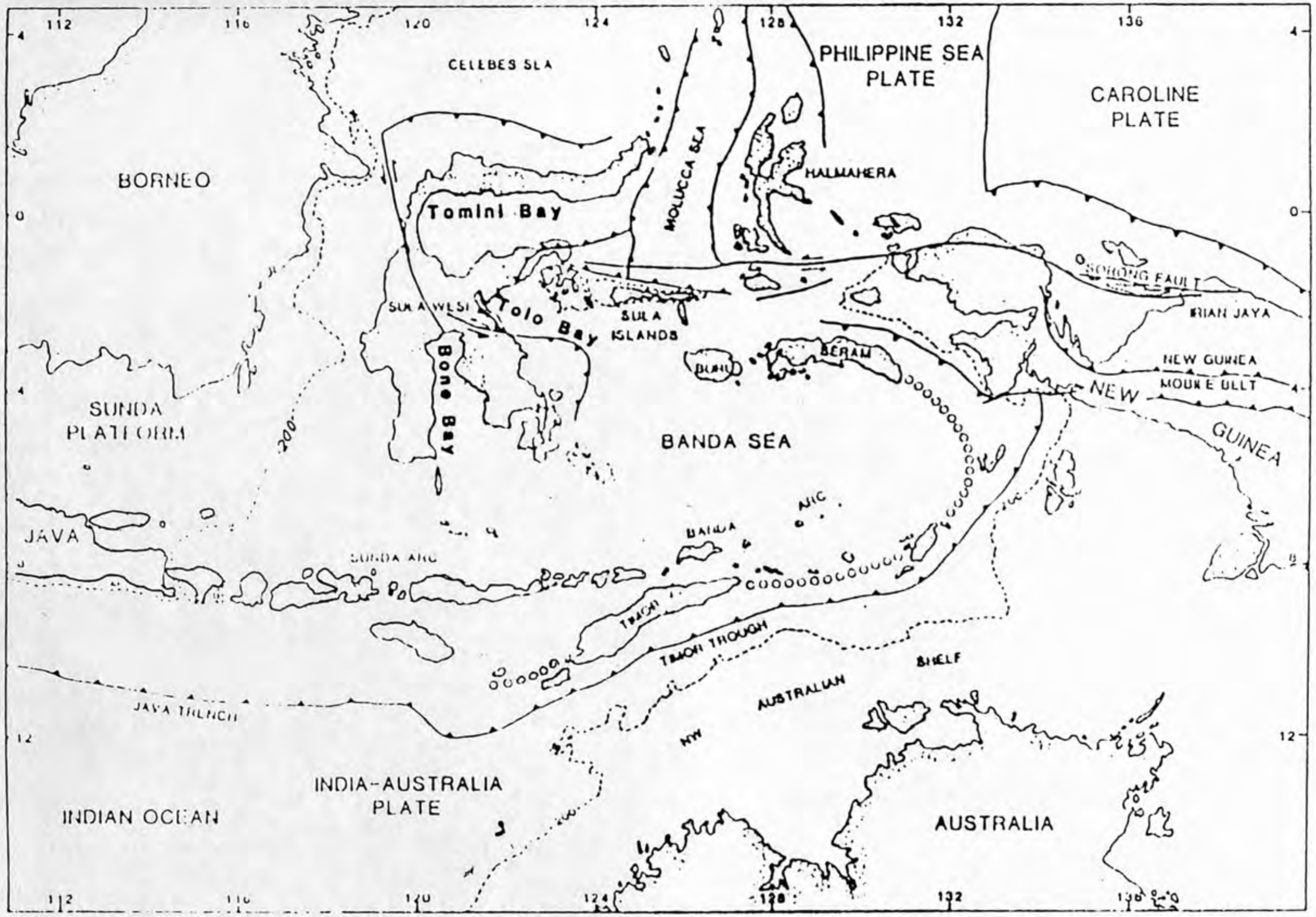


Fig. 5.5A Map showing tectonic configuration of eastern Indonesia (After Hamilton, 1979; Silver, 1979) and the area studied. Teeth along thrusts are on the overriding plate.

Fig. 5.5 Diagram showing the stratigraphy of the East Arm of Sulawesi and the Papua New Guinea and East Indonesia continental fragments, including Banggai-Sula, Tukang Besi-Buton, Buru, Seram, Misool with Irian Jaya microcontinents (platforms). It is apparent that the three episodes of tectonic divergence might have occurred in Triassic-Jurassic, Late Jurassic-Early Cretaceous and Palaeocene times.

<u>Sources of the stratigraphy:</u>

1. Banggai-Sula Island : Sato et al., 1978; Westermann et al., 1978; Surono & Sukarna, 1984; Supanjono et al, 1984.

2. Buton : Suharsono, 1973; Sikumbang, 1983; Smith, 1983; Koswara & Sukarna, in prepar.

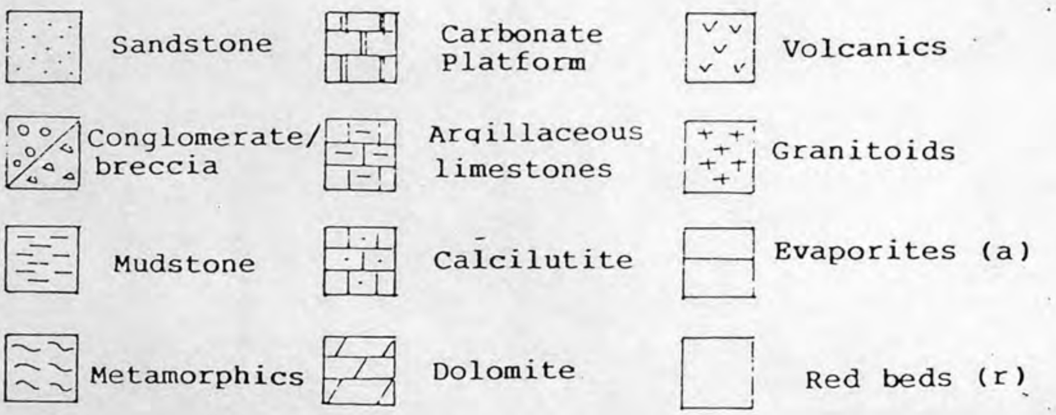
3. Buru : Tjokrosaputro & Budhitrisna, 1981.

4. Seram : Audley Charles et al., 1979; Kastowo, 1982.

5. Misool : Rusmana et al., 1982.

6. Irian Jaya : Visser & Hermes, 1962; Hermes, 1974; Pigram & Panggabean, 1981; Pigram & Sukamta, 1982; Robinson et al., 1982a,b; Pieters et al., 1982, 1983; Dow et al., 1984.

7. Papua New Guinea : Dow et al., 1984; Pigram & Panggabean, 1984.



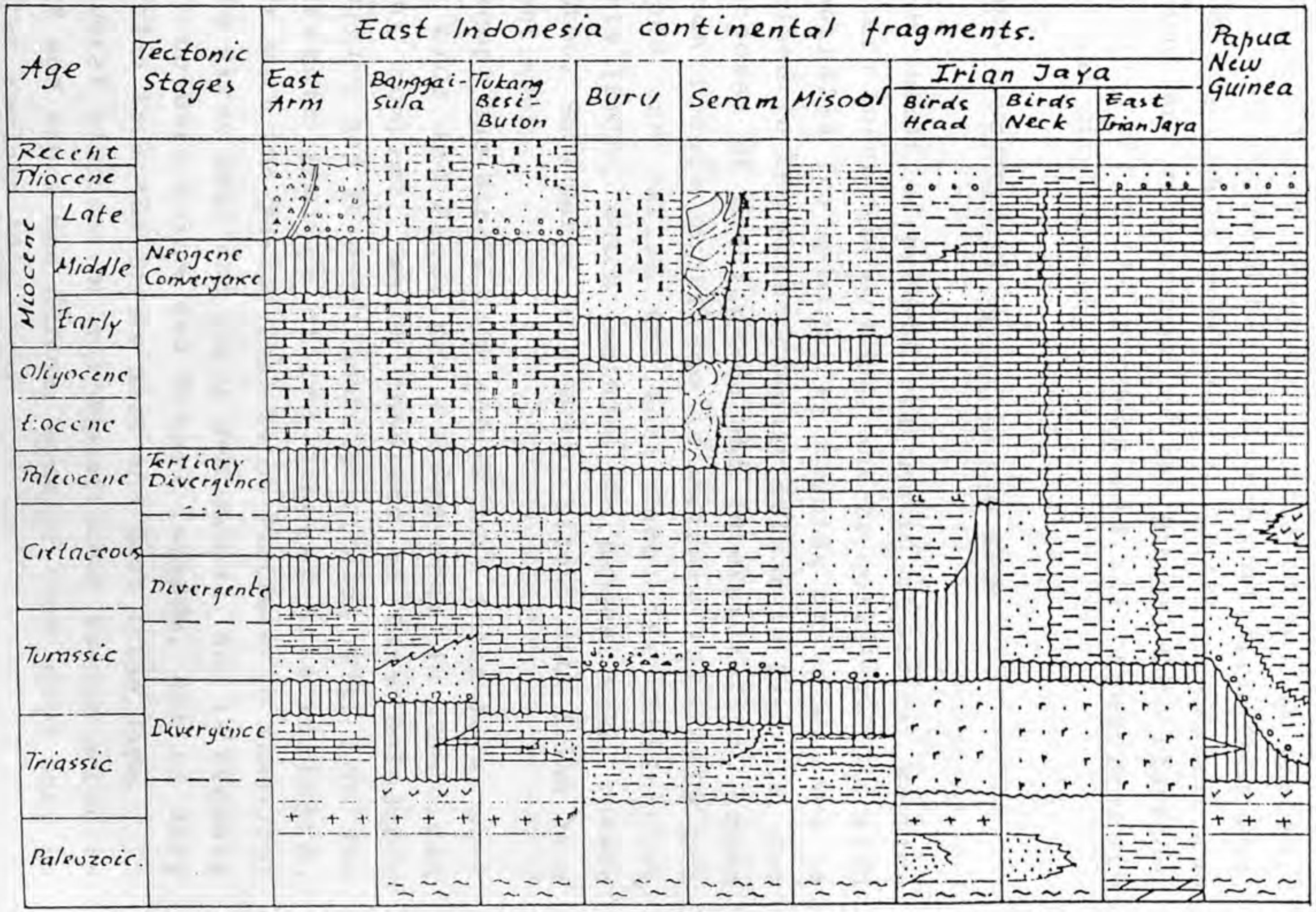

Age
Tectonic Stages
East Indonesia continental fragments.
East Arm
Banggai-Sula
Tukang Besi-Buton
Buru
Seram
Misool
Irian Jaya
Birds Head
Birds Neck
East Irian Jaya
Papua New Guinea
Recent
Pliocene
Miocene
Late
Middle
Early
Oligocene
Eocene
Paleocene
Cretaceous
Jurassic
Triassic
Paleozoic
Neogene Convergence
Tertiary Divergence
Divergence
Divergence

Fig. 5.5

5.4 TECTONIC ORIGIN OF THE BANGGAI-SULA PLATFORM

5.4.1 Introduction

Kinematic models for tectonic origin of the Banggai-Sula Platform are essentially related to the tectonic divergence of northern margin of the Australian Continent. Two extreme models have been forwarded to explain the detachment and displacement of the Banggai-Sula Platfrom, namely i) rift-drift process and ii) trunscurrent-transformal displacement.

5.4.2 Non-depositional events in the EICF

The result of present systematic geological mapping of eastern Indonesia by GRDC, and by collaboration with BMR Australia for geological mapping of Irian Jaya is summarised in the stratigraphic column in Fig. 5.5. For the purposes of description and interpretation, those microcontinents occurring in eastern Indonesia are regrouped and called the East Indonesia Continental Fragments (EICF), which includes the Banggai-Sula, Tukang Besi-Buton, Buru-Seram, Obi-Bacan, Misool with Irian Jaya microcontinents.

Except the Bird's Neck and Eastern Irian Jaya, most of the EICF and Papua New Guinea show an identical basement complex, which consists of Late Paleozoic metamorphics and Permo-Triassic granitoids (Fig. 5.5). The pre-Jurassic geology of the Banggai-Sula Platform, together with the Tukang Besi-Buton Platform is identical with the Birds Head and Papua New Guinea, and is closely similar to Buru-Seram and Misool Islands, but is very distinctive from the Bird's Neck and East Irian Jaya.

Several hiatuses or non-depositional events occur in the EICF and Papua New Guinea, some of them occur only in particular regions (Fig. 5.5):

(i) **Early Jurassic hiatus** occurs throughout the EICF and Papua New Guinea.

(ii) **Early Cretaceous hiatus** occurs only in Banggai-Sula and Tukang Besi-Buton Platforms.

(iii) **Paleocene hiatus** occurs in the Banggai-Sula, Tukang Besi-Buton and Buru-Seram microcontinents.

(iv) **Middle Miocene hiatus** occurs only in the Banggai-Sula (i.e. in the East Arm of Sulawesi) and Tukang Besi-Buton Platforms.

(i) Jurassic Unconformity

The Jurassic hiatus occurs everywhere in the Eastern Indonesia Continental Fragments (EICF) and Papua New Guinea (Fig. 5.5). This hiatus is interpreted as post-breakup stage (i.e.unconformity) of rift-drift sequence in northern margin of the Australian Continent (Pigram & Panggabean, 1984).

In the East Arm of Sulawesi, the hiatus is marked by the occurrence of the Kapali Beds (Fig. 5.2). This unit is similar to the Kabau Formation and Bobong Formation in Banggai-Sula Islands, which includes a conglomeratic basal succession with clasts derived from the Banggai-Sula basement complex (Surono & Sukarna, 1985; Supanjono & Haryono, 1985).

The basinal setting shows a significant deepening toward the west in Late Jurassic time. Since then, sedimentation has continued uninterrupted until the Recent in Papua New Guinea, up to Late Eocene in Misool and up to the end of Cretaceous in Buru-Seram, but only up to Late Jurassic in Banggai-Sula and Tukang Besi-Buton Platforms.

The Jurassic unconformity is interpreted to have occurred everywhere in the world, due to eustatic falls of sea-level (Fig. 5.7, Vail, et al., 1977).

(ii) Early Cretaceous Unconformity

Early Cretaceous deposits do not occur in the Banggai-Sula and Tukang Besi-Buton Platforms (Fig. 5.2). While in the other parts of the EICF and Papua New Guinea continuous sedimentation took place (Fig. 5.3). In the Banggai-Sula and Tukang Besi-Buton Platforms, the Jurassic carbonate shelf sequence is overlain paraconformably by Late Cretaceous bathyal calcilutite (i.e. the Luok Beds and Tanamu Formation in the East Arm of Sulawesi and Banggai-Sula Islands, respectively, Fig. 5.2, 5.5). No coarse clastic rocks are found at the base of the calcilutite. These features suggest the Early Cretaceous non-depositional event in the Banggai-Sula and Tukang Besi-Buton Platforms is a submarine hiatus resulting from local sea-level changes coupled with or superimposed the tectonic divergence dominated by transcurrent-transformal displacement of the platforms from the EICF.

The occurrence of a submarine hiatus is related to rapid rise of sea-level, with sediment starvation in the deep offshore because of temporary entrapment of detritus in coastal delta complexes (Miall, 1986). Vaïl et al. (1984) stated that the age of a submarine hiatus wihin a given depositional sequence tends to be synchronous globally, but may differ slightly from basin to basin with changes in rate of deposition and subsidence.

The occurrence of an Early Cretaceous submarine hiatus in the Banggai-Sula and Tukang Besi-Buton Platforms is interpreted as indicating the uplift of these regions due to tectonic divergence, with a rise in sea-level in the outer shelf, in Late Cretaceous. Following the hiatus calcilutite-dominated sediments were deposited in bathyal depth paraconformably on top of the Jurassic sediments (Fig. 5.2,).

The absence of clastic rocks at this unconformity is compatible with the temporary entrapment of detritus in coastal delta complex (Miall, 1986).

Vail et al.(1972) show an abrupt fall of sea-level in Early Cretaceous for a short period, which was followed by a rapid rise sea-level in the Late Cretaceous for a short period (Fig. 5.7). This feature might be related to the development of restricted or local hiatus, which is due to or superimposed tectonic activity (i.e. transcurrent-transformal displacement of the BSP from the EICF).

(iii) <u>Palaeocene Unconformity</u>

A Paleocene hiatus occurs in the Banggai-Sula, Tukang Besi-Buton and Buru-Seram microcontinents. The hiatus indicates the uplifting of these regions subsequent to transformal displacement from the EICF. During this sea-level fall, the shelf underwent erosion, and hence, there is no sedimentary record in these microcontinents (Fig. 5.3). In the Banggai-Sula Islands, the unconformity is marked by the occurrence of basal coarse clastic rocks at the base of the Palaeogene carbonate platform (Surono & Sukarna, 1985; Supanjono & Haryono, 1985). In Irian Jaya, Misool and Papua New Guinea, however, carbonate platform and subsidiary clastics were deposited continuously in neritic depth (Fig. 5.5).

Vail et al.(1972) show slight fall of sea-level in Palaeocene time, which might be related to local unconformity (Fig. 5.7).

(iv) <u>Middle Miocene Unconformity</u>

The Middle Miocene hiatus is related to tectonic convergence in eastern Sulawesi, in which the Banggai-Sula and Tukang Besi-Buton Platforms collided with the Eastern Sulawesi Ophiolite Belt (EOSB). The hiatus is marked by the occurrence of coarse clastic rocks (Batui Group) deposited on top of the collision complex. This will be discussed later in this chapter.

5.4.3 **Rift-drift process.**

The detachment was initiated by rifting (breakup) of the continental margins and followed by transcurrent-transformal displacement of the continental fragments. Pigram and Panggabean (1984) following Falvey and Mutter (1982), on the basis of recognition of rift-drift sequence in eastern Indonesia, suggest that the Banggai-Sula Platform was detached from a region in Central Papua New Guinea, about 141° E - 145° E longitude. They speculate, further, that the oceanic crust fragment during Triassic rifting is no older than Early Jurassic in the east and Middle Jurassic in the west adjacent to the newly formed continental margin of Papua New Guinea, which marks the separation of the EICF from the Australain Continent.

The tectonic divergence marks the overall and regional rifting (breakup) of the northern margin of the Australian Continent, which began in the Permian (230 Ma), and gave rise to the detachment of the East Indonesia Continental Fragments (EICF) from the Papua New Guinea and Australian Continent.

Schneider (1972) and Falvey (1974) pointed out that rift-drift process consists of three distinct stages of tectonic event: pre-breakup, breakup and post-breakup. Each stage shows a characteristic association of sedimentary processes and related tectonic events. Falvey and Mutter (1981) described the characteristics of these stages for the margin of the Australian Continent.

The pre-breakup stage is marked by a long period (50 Ma or more) of rift valley tectonism and a high rate of accumulation of both marine and continental sediments.

The breakup stage is characterised by major faulting, uplift giving rise to fall of sea-level, erosion and local volcanism. The breakup episode corresponds to the first creation of new oceanic crust subsequent to tectonic divergence of continental crust, and is defined as the initation of seafloor spreading.

The post-breakup stage is marked by subsidence of continental margin and the development of initial restricted marine sedimentation, following by open marine deposition with relatively low rates of accumulation. The basal succession of these sediments is marked by the post-breakup unconformity, which may be a disconformity, paraconformity or a transgressive or regressive cycle.

The Early Mesozoic tectonic divergence in the EICF and Papua New Guinea contains these three stages of the rift-drift sequences :

The pre-breakup stage is characterised by the occurrence of a Carboniferous or older metamorphic basement complex in the EICF and Papua New Guinea, except the East Irian Jaya, which is overlain by shallow marine to paralic sediments deposited during pre-breakup stage (Pigram and Panggabean, 1984). In the Banggai-Sula, Tukang Besi-Buton, Obi-Bacan and Papua New Guinea microcontinents the metamorphic complex is intruded by Permo-Triassic granitoids, which are overlain by Triassic volcanics (Sukamto, 1975; Hamilton, 1979; Audley Charles et al., 1979; Azwan, 1981; Pigram & Panggabean, 1983; Dow et al., 1984; Surono & Sukarna; 1985; Supanjono & Haryono, 1985; Koswara & Sukarna, 1986; Rusmana et al., 1986). In the east Arm of Sulawesi, a granitoid basement complex occurs at about 800 m offshore of Mentawa in Tolo Bay.

The breakup stage is marked by the accumulation of clastic and carbonate sediments ranging from paralic to slope deposits. The breakup sequence in Papua New Guinea and Banggai-Sula and Tukang Besi-Buton Platforms is closely similar (Fig. 5.5).

The post-breakup stage is marked by a regional hiatus everywhere in Papua New Guinea and EICF. The initial basal succession is characterised by the deposition of coarse clastics of paralic to shallow marine followed by shelf carbonates. The basinal setting shows a significant deepening toward the west in Late Jurassic time. Since then, sedimentation has continued uninterrupted until the Recent in Papua New Guinea, up to Late Eocene in Misool and up to the end of Cretaceous in Buru-Seram, but only up to Late Jurassic in the Banggai-Sula and Tukang Besi-Buton Platforms.

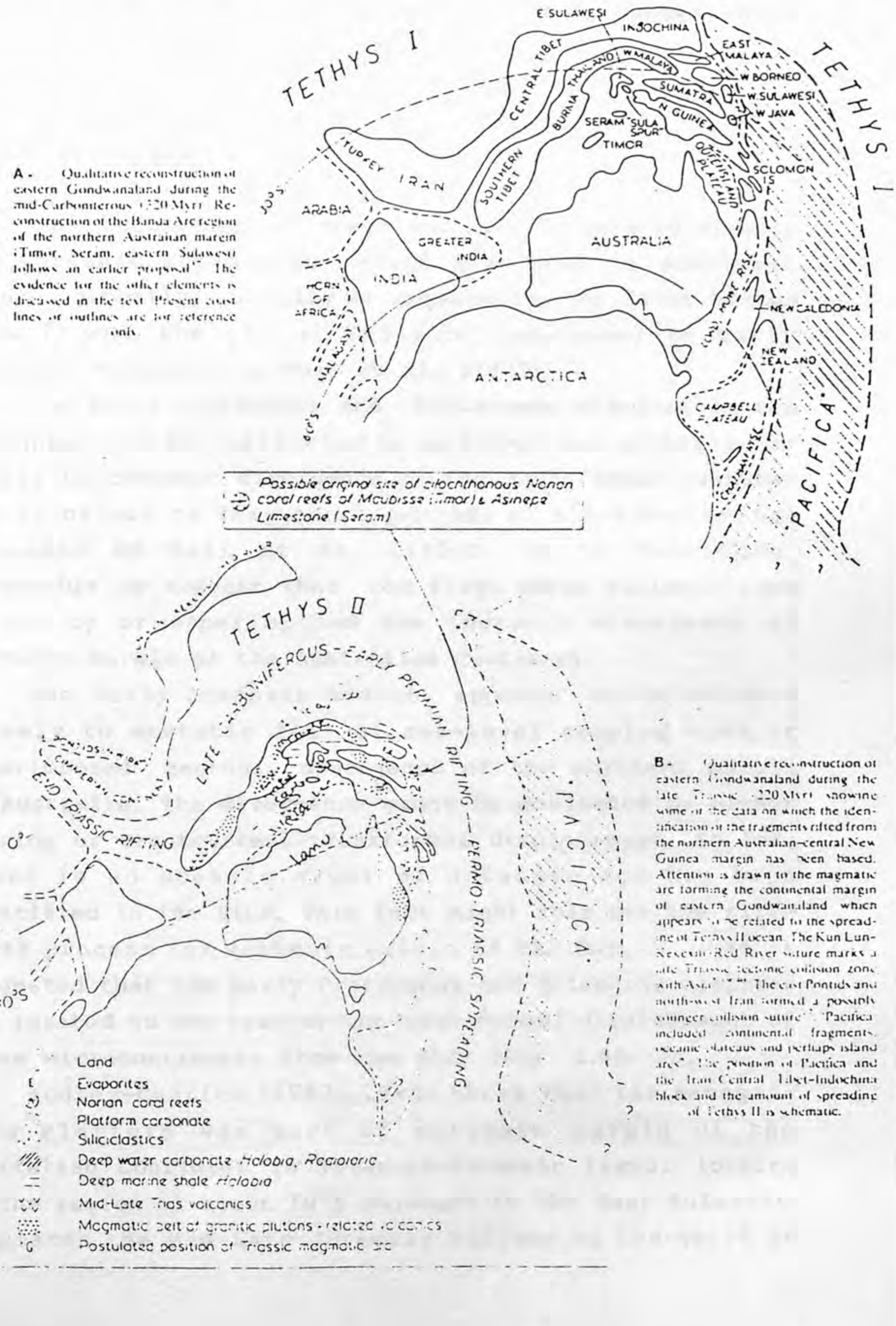
The rift-drift sequence has been recognised along the former northern margin of the Australian Continent and the age of the post-breakup unconformity in the Papua New Guinea and EICF correlates with the commencement of seafloor spreading. Pigram and Panggabean (1984) suggest that oceanic crust formed during this tectonic divergence, is no older than Early Jurassic in the east and Middle Jurassic in the west adjacent to the newly formed continental margin of Papua New Guinea, marking the separation of the EICF from the Australian Continent.

The main geological constraint relating to rift-drift process is that the Jurassic oceanic crust or ophiolitic rocks which were formed during post-Triassic rifting, has so far, not been identified in eastern Indonesia. In accepting or ruling out this model, it is necessary to have definite ages for the ophiolitic rocks occurring in eastern Indonesia.

5.4.4 Transcurrent-transformal displacement

The Banggai-Sula Platform is generally regarded as having been detached from western Irian (New Guinea) in Late Cainozoic and displaced westward along the Sorong Fault Zone (Visser & Hermes, 1962; Krause, 1965; Hermes, 1968, 1974; Gribi, 1973; Hamilton, 1978, 1979; Katili, 1978; Norvick, 1979; Silver and Smith, 1983; Smith, 1983), or north of Misool region (Audley-Charles et al., 1972;

Fig. 5.6.1 Map showing qualitative reconstruction of eastern Gondwanaland and inferred position of the Banggai-Sula Platform in the Misool region, about 45° S latitude during Mid-Carboniferous (A) and about 12° S latitude in Late Triassic (B), (After Audley-Charles, 1983).



A - Qualitative reconstruction of eastern Gondwanaland during the mid-Carboniferous (320 Myr). Reconstruction of the Banda Arc region of the northern Australian margin (Timor, Seram, eastern Sulawesi) follows an earlier proposal. The evidence for the other elements is discussed in the text. Present coast lines or outlines are for reference only.

B - Qualitative reconstruction of eastern Gondwanaland during the late Triassic (220 Myr) showing some of the data on which the identification of the fragments rifted from the northern Australian–central New Guinea margin has been based. Attention is drawn to the magmatic arc forming the continental margin of eastern Gondwanaland which appears to be related to the spreading of Tethys II ocean. The Kun Lun-Nexexin-Red River suture marks a late Triassic tectonic collision zone with Asia. The east Pontids and north-west Iran formed a possibly younger collision suture. 'Pacifica' included continental fragments, oceanic plateaus and perhaps island arcs. The position of Pacifica and the Iran-Central Tibet-Indochina block and the amount of spreading of Tethys II is schematic.

Audley-Charles, 1983, 1984). This assumption is based on correlation between the geology of the Banggai-Sula Islands and western Irian Jaya.

5.4.5 Discussion

The occurrence of these hiatuses is related closely to tectonic activities, which give rise to sea-level changes globally, locally or regionally. The first hiatus fits in with the global falls of sea-level in Early Jurassic suggested by Vail et al. (1972).

The Early Cretaceous and Palaeocene hiatuses in the BSP appear to be restricted unconformities probably due mainly to tectonic divergence of the EICF. These hiatuses are identical to very short periods of sea-level change proposed by Vail et al. (1972). It is therefore, reasonable to suggest that the first three hiatuses were caused by or superimposed the tectonic divergence of northern margin of the Australian Continent.

The Early Jurassic hiatus, appears to be related closely to eustatic fall of sea-level coupled with or superimposed tectonic divergence of the northern margin of Australia. The divergence might be dominated by either rifting or transcurrent-transformal displacement. So far, there is no oceanic crust of Jurassic age has been identified in the EICF. This fact might rule out the rift-drift process for tectonic origin of the BSP. It is suggested that the Early Cretaceous and Paleocene hiatuses are related to the trancurrent-transformal displacement of these microcontinents from the EICF (Fig. 5.8B, C).

Audley-Charles (1983, 1984) shows that the Banggai-Sula Platform was part of northern margin of the Australian Continent in Triassic-Jurassic times, located in the region of about 10°S adjacent to the East Sulawesi. He places the Mid-Late Jurassic rifting to the north of

Fig. 5.6.2 Map showing a qualitative reconstruction of eastern Gondwanaland during Carboniferous time, showing position of the BSP and ESOB in Misool region at about 35° S latitude. (After Audley-Charles, 1984).

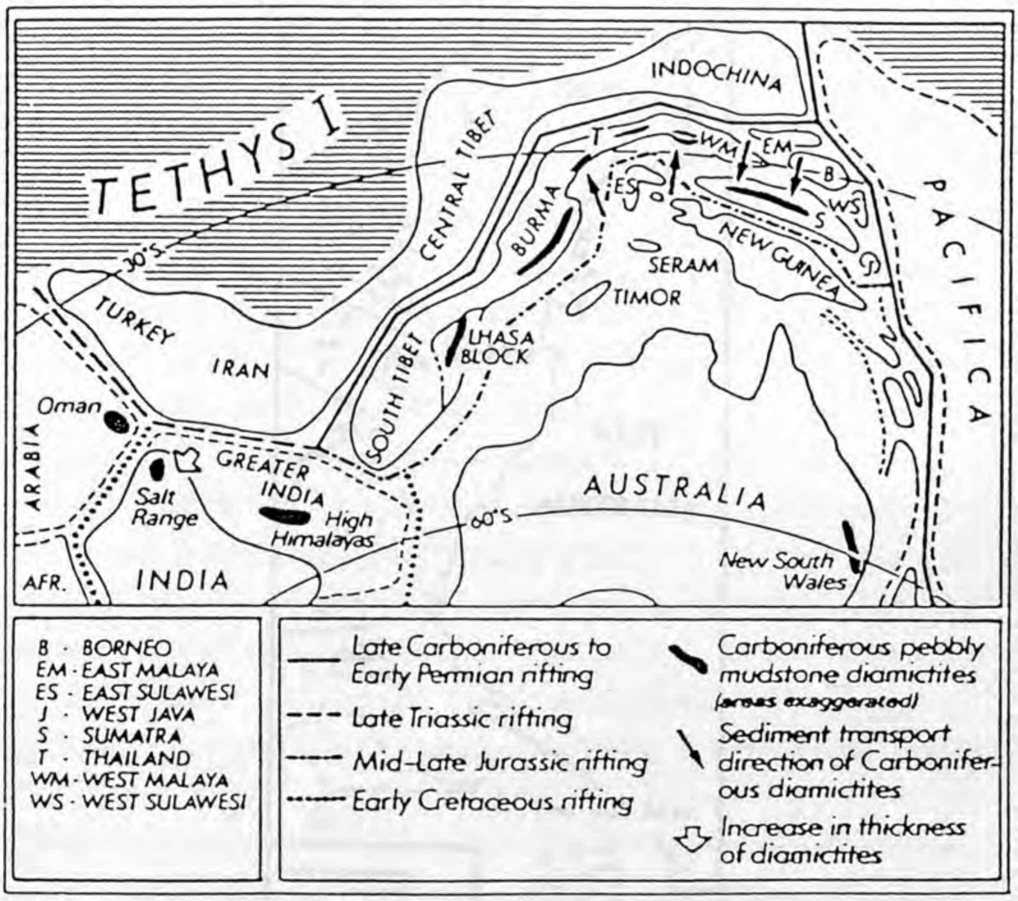


Fig. 1 Correlation of the late Carboniferous marine glacial pebbly mudstones (diamictites) of the eastern part of greater Gondwanaland. The qualitative reconstruction[2] is based on the Antarctic–India–Australia assembly of ref. 4. Present coastlines or outlines are for reference only. The shapes of the Asian fragments originating in greater Gondwanaland have been modified by tectonic deformation since they left Gondwanaland. The original shapes have not been reconstructed here but some allowance for crustal shortening has been made. Late Carboniferous marine glacial diamictite data for Sumatra, Malaya, Thailand, Burma, New South Wales, High Himalayas are from ref. 3. Oman glacial marine diamictites are plotted from ref. 8. The western Salt Range diamictites described by Teichert[9] may be early Permian rather than late Carboniferous. Location of Pacifica and late Carboniferous to Permian and late Triassic rifting are after ref. 2, Jurassic rifting after ref. 5. Early Cretaceous rifting and latitudes are plotted after ref. 4.

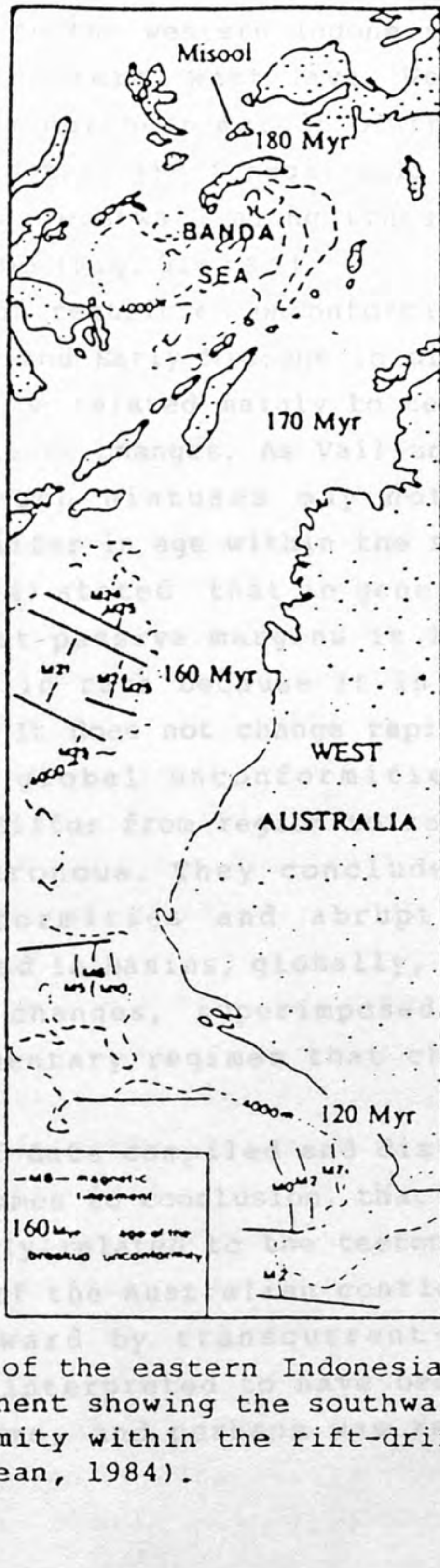


Fig. 5.6.3 Map of the eastern Indonesia and western Australian continent showing the southward younging of the breakup unconformity within the rift-drift sequence (After Pigram & Panggabean, 1984).

the Irian, during which the western Indonesia continental fragments (include Sumatera, West Java, West Sulawesi, Borneo) detached from northern margin of the Australian Continent. He suggests that the Banggai-Sula Platform was detached and displaced westward along transcurrent fault in Late Cainozoic time (Fig. 5.6.1 & 2)

The occurrence of restricted unconformities in Early Cretaceous, Paleocene and Early Miocene in different parts of the EICF seems to be related mainly to tectonics which caused regional sea-level changes. As Vail and Todd (1981) pointed out that such hiatuses may not be global, sometimes they may differ in age within the same basin.

Vail et al. (1984) stated that in general, tectonic subsidence along most passive margins is long term and gradually decreases in rate because it is related to a thermal decay curve. It does not change rapidly enough to cause regional or global unconformities. Tectonic subsidence patterns differ from region to region, and are not globally synchronous. They conclude, that many synchronous unconformities and abrupt changes in stratigraphy observed in basins, globally, are caused by eustatic sea-level changes, superimposed on regional tectonics and sedimentary regimes that change at much slower rates.

On the basis of data compiled and discussed above, the present author comes to conclusion, that the origin of the BSP is essentially related to the tectonic divergence of northern margin of the Australian continent, and the displacement westward by transcurrent-transformal movement, which is interpreted to have been started in Early Cretaceous time, and perhaps was reactivated in Palaeocene time.

5.5 IMPLICATION FOR THE AGE OF THE BANDA SEA

Katili (1974) put forward the hypothesis that the Banda Sea represents old oceanic crust. Hamilton (1979) and Carter et al. (1976) are of the opinion that the Banda Sea crust is Late Tertiary in age. Bowin et al. (1980) favoured a Cretaceous or older age for the Banda Basin, and considered that the basin is floored by a trapped piece of oceanic crust.

Lapouille (1985) on the basis of magnetic studies, concluded that the Banda Sea crust had been formed in Early Cretaceous time as part of the Western Pacific Ocean and Eastern Indian Ocean and was trapped in Miocene time. He also correlates the Banda Sea with the Argo Abyssal Plain in the northeast of Indian Ocean. He points out that the North Banda Sea crust is younger that the South Banda Sea crust, and was created in Middle to Late Cretaceous.

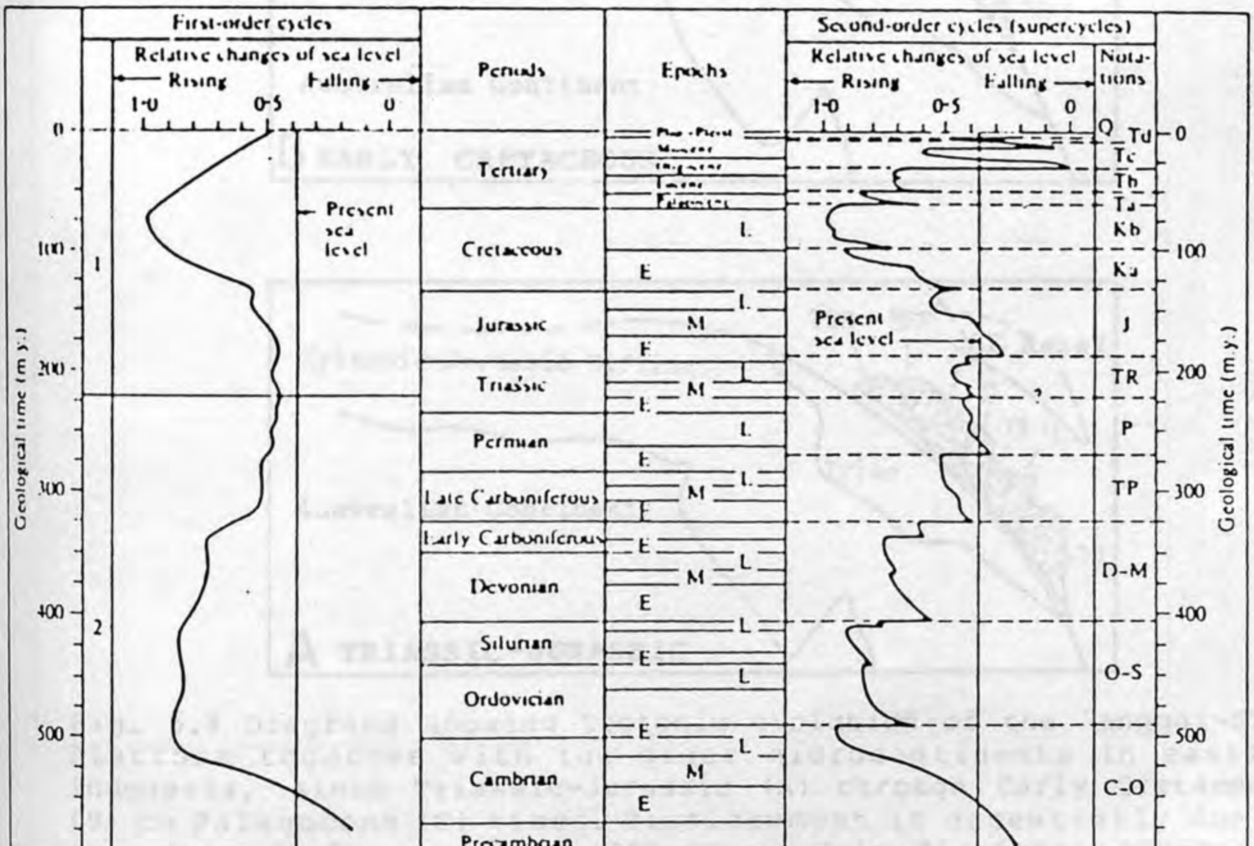
Chao and McCabe (in press) based on identification of magnetic reversal ages, heatflow analysis, bathymetry and on-land geology suggest that the Banda Sea was formed in the Cretaceous. They speculate, that the Banda Sea, Celebes Sea, and Sulu Sea formed a continuous ocean basin during Cretaceous to Early Tertiary times.

The ophiolitic rocks in the East Arm of Sulawesi range in age from Early to Late Cretaceous with Eocene to Early Oligocene seamounts. The older age of the Balantak Ophiolite is similar to the age of the Banda Sea crust. Hence, the Balantak Ophiolite is considered as part of the Banda Sea Crust, the leading wedge of which was subducted beneath the continental margin and/or volcanic arcs of the Sundaland in Late Cretaceous to Paleocene times.

Pigram and Panggabean (1983) suggested that the Banda Sea was floored by Late Jurassic to Cretaceous oceanic crust. According to Pigram and Panggabean (1983) Jurassic ophiolitic rocks occur in association with the continental margins of the East Indonesia Continental Fragments

(EICF), but so far, no ophiolitic rocks of Jurassic or older have been identified definitely in the Eastern Sulawesi Ophiolite Belt (ESOB). Pre-Cretaceous ophiolitic rocks are present in the Argo Abyssal Plain but along the Australian Continental Margin further east have been subducted beneath the Banda Arcs and beneath the Palaeogene Volcanic Arc which collided with northern New Guinea in mid-Miocene times.

Fig. 5.7 Diagram showing the Phanerozoic eustatic curves in terms of first- and second-order cycles (After Vail et al., 1977).



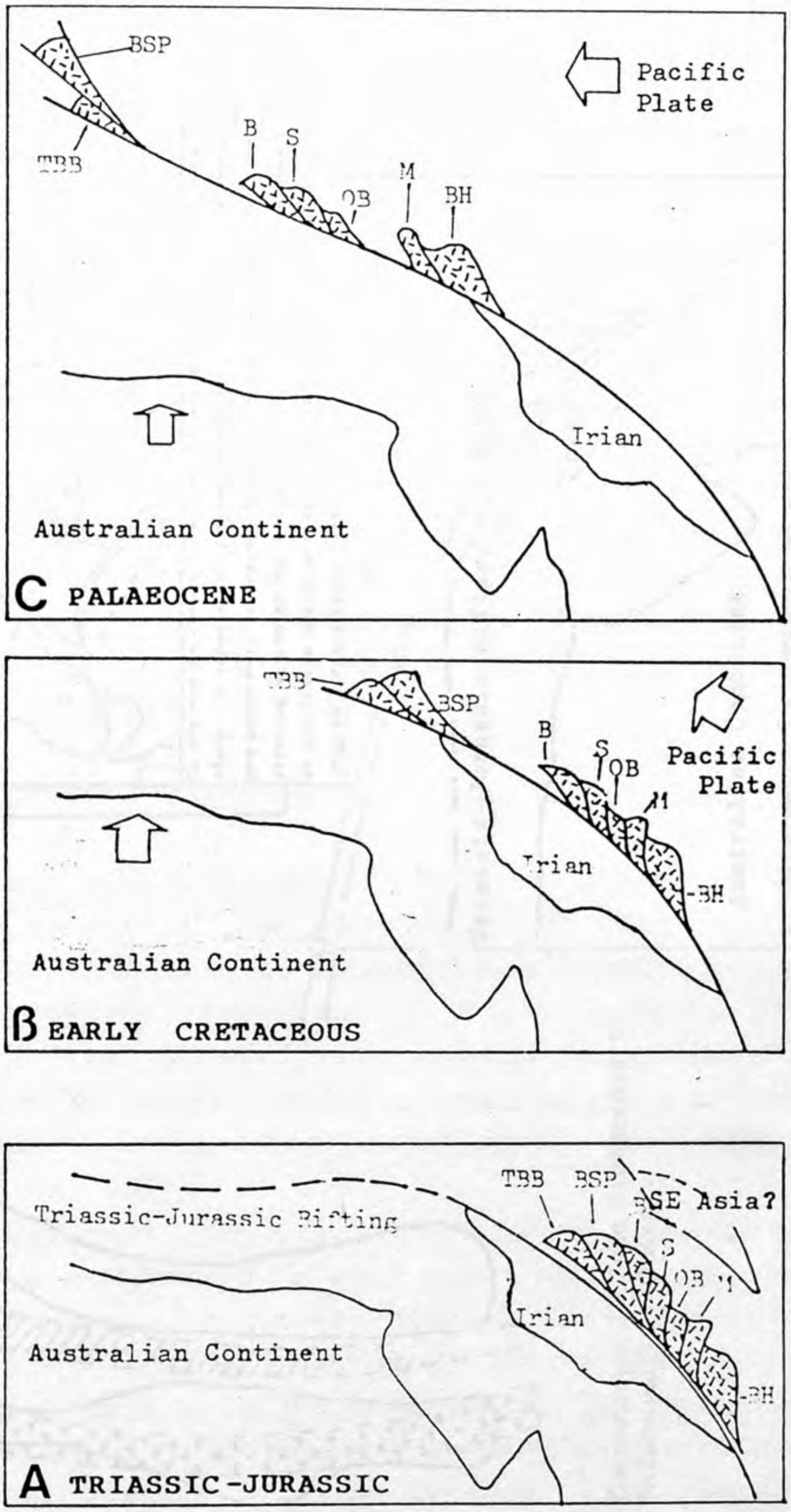


Fig. 5.8 Diagrams showing tectonic evolution of the Banggai-Sula Platform together with the other microcontinents in eastern Indonesia, since Triassic-Jurassic (A) through Early Cretaceous (B) to Palaeocene (C) times. Displacement is essentially due to transform fault movement. BSP=Banggai-Sula Platform, TBB=Tukang Besi-Buton Platform, B=Buru, S=Seram, OB=Obi-Bacan, M=Misool, BH=Bird's Head microcontinents.

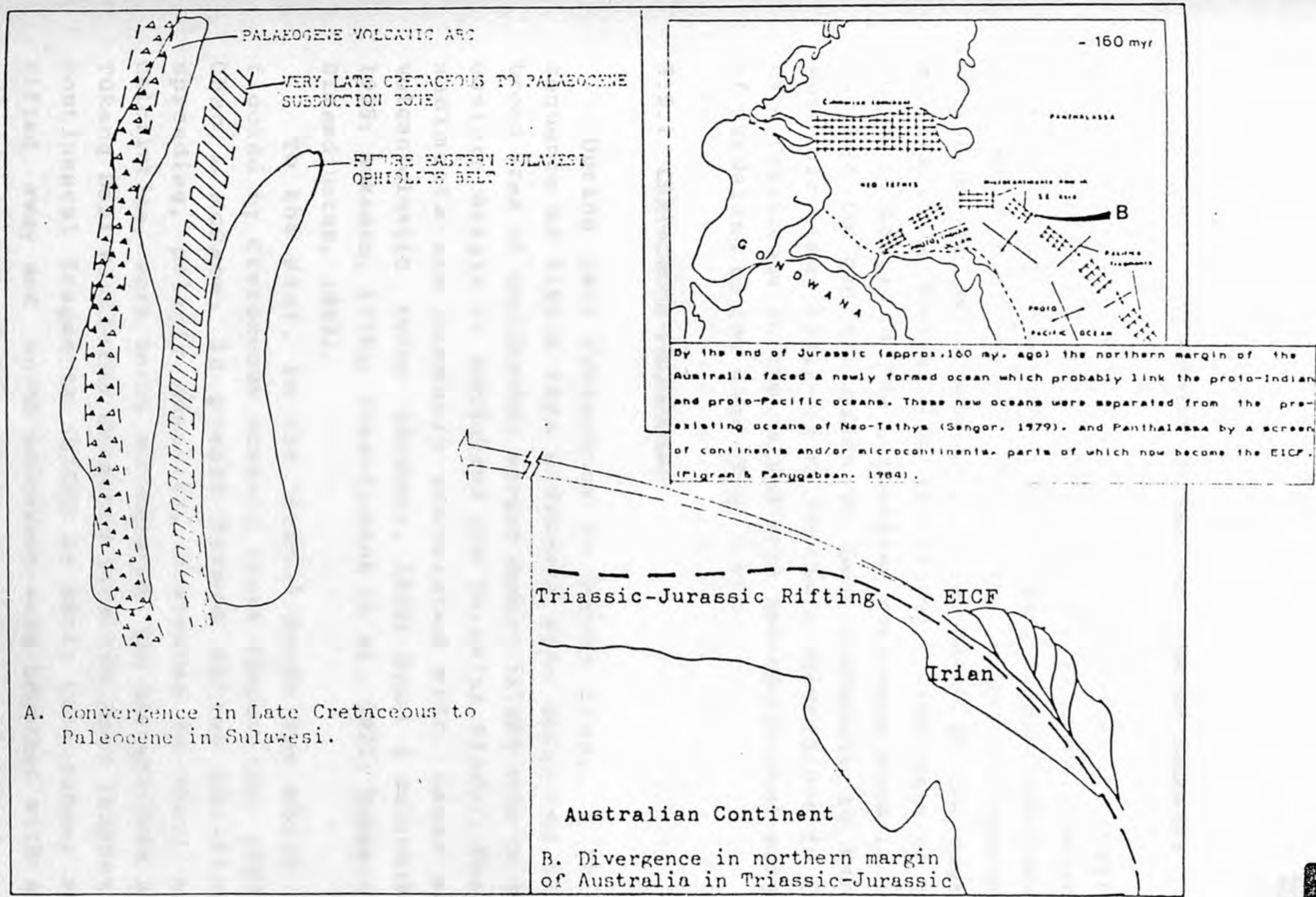


Fig.5.9 map showing tectonic evolution of the Banggai-Sula Platform, which was due to the tectonic divergence from the northern margin of the Australian Continent in Triassic-Jurassic time followed by tectonic convergence in Late Cretaceous-Palaeocene in Sulawesi region

5.6 TECTONIC DEVELOPMENT OF THE EAST ARM OF SULAWESI

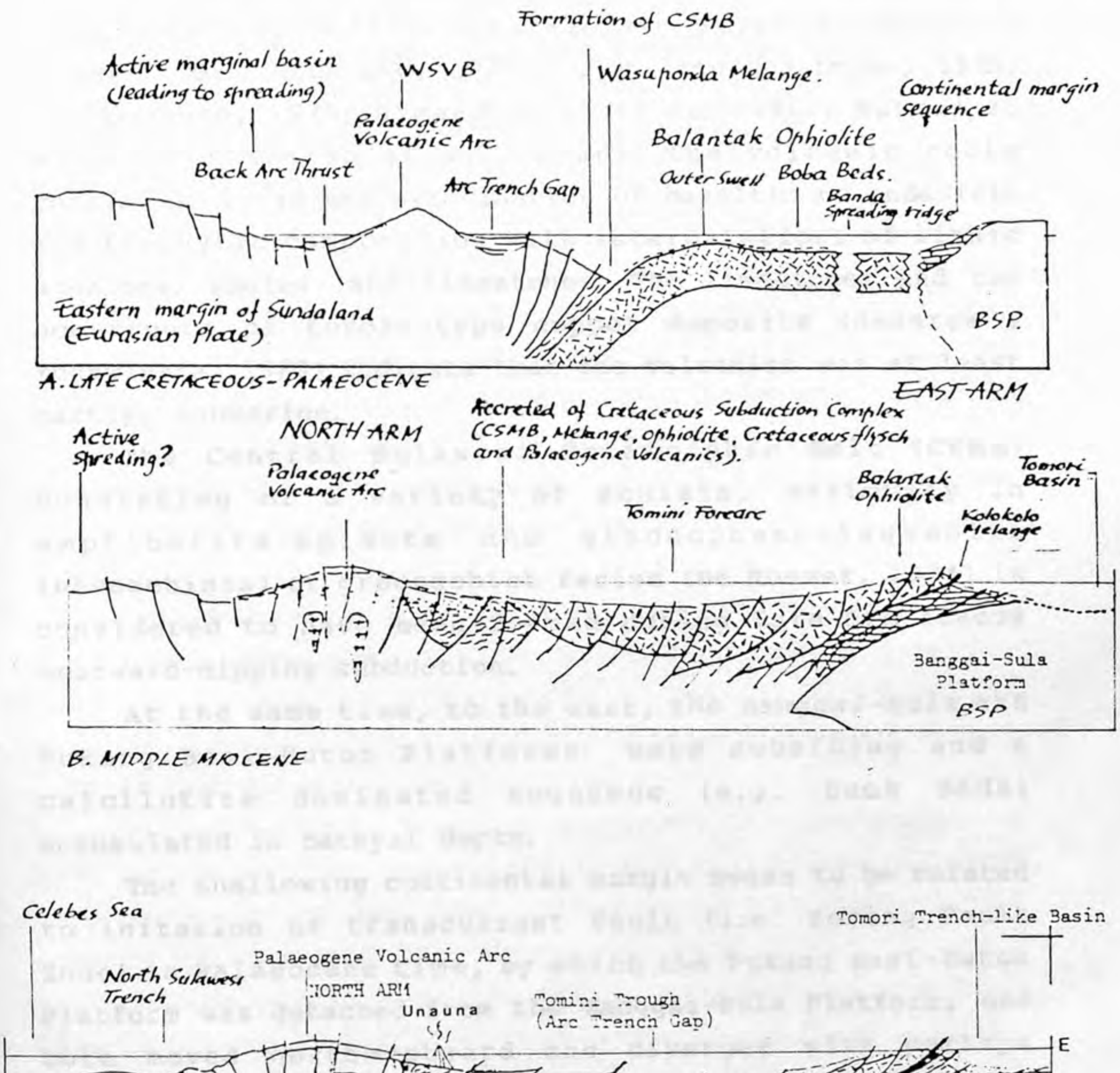
The tectonic evolution of the East Arm as an integral part of Sulawesi and its surroundings, is closely related to the tectonic development of the Banggai-Sula and Tukang Besi-Buton Platforms. The kinematic model for tectonic evolution and palaeogeographic development of the region appears to be dominated by the tectonic divergence of the northern margin of the Australian continent when it lay further to the southeast in Late Paleozoic to Early Mesozoic time, followed by tectonic convergence in the Late Cretaceous in the island arcs and continental margin of Sundaland to the west (Fig. 5.9).

5.5.1 CRETACEOUS-PALAEOGENE

During Late Cretaceous to Eocene time, a thick sequence of flysch type sediments were deposited in a broad area of continental margin and/or island arcs or the eastern margin of Sundaland (SE Eurasian Plate). These sediments are commonly associated with lavas and volcaniclastic rocks (Brouwer, 1934; Djuri & Sujatmiko, 1975; Sukamto, 1975a; Simandjuntak et al., 1981; Sukamto & Simandjuntak, 1982).

To the east, in the 'proto' Banda Sea which is floored by Cretaceous oceanic crust (Lapouille, 1985), Chao & McCabe, in press) formed during sea-floor spreading, pelagic deposits, dominated by chert and calcilutite, were being accumulated. The Banggai-Sula and Tukang Besi Platforms detached from the East Indonesia continental fragments (EICF) in Early Cretaceous, and rifted away and moved northwest-ward together with the oceanic crust of the Banda Sea, part of which was subducted in Late Cretaceous times, beneath the island arcs and continental rise of Sundaland (Fig. 5.10A). The subduction zone dipped westward, and now is marked by the

Fig 5.10 Palinpastic sections showing tectonic evolution of the East Arm of Sulawesi.



Wasuponda Melange in Central Sulawesi (Simandjuntak, 1980).

This subduction zone generated a Palaeogene volcanic arc (Fig. 5.2, 5.10A), the Western Sulawesi Volcano-Plutonic Belt (Sukamto, 1975a; Djuri and Sujatmiko, 1975; Van Leeuwen, 1979; Simandjuntak et al., 1981; Ratman et al., 1986; Sukido et al., 1986). The volcanic rocks consist of lavas and pyroclastics of basaltic to andesitic and trachytic composition with intercalations of lithic arenites, shales and limestones. The limestones and the occurrence of Kuroko-type copper deposits (Sunarya & Yudawinata, 1980) indicate that the volcanism was at least partly, submarine.

The Central Sulawesi Metamorphic Belt (CSMB) consisting of a variety of schists, variously in amphibolite-epidote and glaucophane-lawsonite (blueschists) or greenschist facies (De Roever, 1974) is considered to have been formed during Late Cretaceous westward-dipping subduction.

At the same time, to the east, the Banggai-Sula and Tukang Besi-Buton Platforms were subsiding and a calcilutite dominated sequence (e.g. Luok Beds) accumulated in bathyal depth.

The shallowing continental margin seems to be related to initation of transcurrent fault (i.e. Sorong Fault Zone) in Palaeocene time, by which the Tukang Besi-Buton Platform was detached from the Banggai-Sula Platform, and both moved northwestward and diverged with perhaps slightly different acceleration; the Banggai-Sula Platform moving NNW-ward and Tukang Besi-Buton moving WNW-ward (Fig. 5.8C).

In the Eocene the continental margins of both Banggai-Sula and Tukang Besi-Buton Platform were shallowing due to local or regional tectonic effects, and the deposition of a carbonate platform sequence (e.g. Salodik Limestones) took place and continued to Early

Miocene time. Formation of the carbonate platform might be also related to build-up of the continental margin sequence (cf. NW shelf Australia) described by Falvey and Mutter (1981).

5.6.2 MIDDLE MIOCENE

In Middle Miocene time, tectonic convergence in the East Arm of Sulawesi resulted in the collision of the continental margin of the Banggai-Sula Platform with the Eastern Sulawesi Ophiolite Belt. A similar tectonic convergence occurred in Buton (Smith, 1983). The initial tectonic convergence is a Tethyan type collision zone, in which the continental margin underplated the ophiolite suite which had been imbricated to form the accretionary prism of the Late Cretaceous subduction complex (Fig. 5.10B).

The underplated continental margin sequence (i.e. Balantak Group) consists of shelf to slope sediments, ranging in age from Triassic to Early Miocene. As described previously in Chapter 2, these sediments are part of the Banggai-Sula Platform.

The overriding Balantak Ophiolite is interpreted to have originated as ocean floor material with isolated seamounts, which had been part of the the underplated oceanic crust during Late Cretaceous subduction.

The ophiolitic rocks in the western part of the East Arm, must be older than the underlying Valanginian radiolarian chert (i.e. the Boba Beds). These rocks appear to be part of the underplated oceanic crust during the Late Cretaceous to Paleocene subduction, but now occur in the imbricated complex of the overriding plate of the Middle Miocene collision.

As described previously in Chapter 4, the collision zone in the East Arm of Sulawesi, is considered to have been trending in a NS direction, and now is marked by the

Batui Thrust-Balantak Fault System, which is in many places associated with the Kolokolo Melange and oil seeps. The melange contains fragments detached from both the underplated continental margin sequence and the overriding ophiolite belt. The calcareous shale and/or marlstone matrix of the melange contains planktonic foraminifera of Middle Miocene to Pliocene age. This age coupled with the age of the youngest sediments (i.e. Eocene to Early Miocene Salodik Limestones) involved within the collision gives a Middle Miocene age for the initial tectonic convergence in the East Arm of Sulawesi. The same age is also suggested by Smith (1983) for the collision of Tukang Besi-Buton platform against the Eastern Sulawesi Ophiolite Belt.

5.6.3 PLIO-PLEISTOCENE TO RECENT

As described previously in Chapter 4, the physiographic and tectonic configuration of the East arm together with the North Arm of Sulawesi, were greatly affected and modified in Pliocene time. The continuously active or reactivated collision zone in the East Arm, together with the North Sulawesi subduction zone, which gives rise to the NW-SE trending compression, has initiated the dextral movement of the Balantak Fault System, along which the Balantak Ophiolite and the overlying Lonsuit Turbidites were displaced eastward for some 150 km, and the sinistral movement of the Toili Fault System, by which the collision zone has been displaced for some 20 km to the south in Kolo Atas area, as well as the dextral movement of the Ampana Fault zone, along which the collision zone has been displaced northward for at least 60 km and the sinistral movement of the Uekuli Fault, by which the collision zone was rotated and displaced for some 70 km from that occurring in the eastern part of Central Sulawesi (Fig. 4.16).

The physiography and distribution of Neogene and Recent Volcanic Arc (i.e. the Minahasa-Unauna Arc), suggest the collision might be active or have been reactivated at the present time. The recent activity of this collision is also suggested by earthquake activity (Mc Caffrey et al., 1982; Effendy, in preparation). The Neogene to Recent Volcanic Arcs occur in an arcuate and convex northwestward and slightly parallel to the collision zone in the East Arm, and is shifted eastward away from the Palaeogene Volcanic Arc (Fig. 4.18B, Fig. 6.10C).

In Late Miocene, magmatism occurred in the North Arm (Sukamto, 1975; Sukamto and Simandjuntak, 1982). Granitoid rocks intruded the Late Cretaceous-Eocene flysch sediments and the Palaeogene volcanic rocks.

Hamilton (1979) suggests the occurrence of an active, southward-dipping subduction zone in Sulawesi Sea, to the north of the North Arm. Katili and Sudradjat (1983) suggest that this subduction was active recently. If this is so, we have a double subduction zone in northern part of Sulawesi. Alternatively, the north Sulawesi subduction zone might have been active until Pliocene time, but now, has become a back-arc thrust during the Neogene to Recent collision of the East Arm of Sulawesi. The occurrence of double subduction zone in northern Sulawesi might be responsible for the present configuration of the region.

The occurrence of at least three terraces of Quaternary coralline reefs in the south coast of the East Arm indicates the rapid uplift of the region.

5.7 SUMMARY AND CONCLUDING REMARKS

The detailed study of sedimentary successions and tectonic history of the ophiolite belt in the East Arm of Sulawesi reveal some interesting features delineated below:

A. Stratigraphy and Tectonostratigraphy

Sedimentary successions in the East Arm of Sulawesi are divided into 4 tectonostratigraphic units :

1. Balantak Group consists of continental margin sequence dominated by carbonates ranging in age from Triassic to Palaeogene.

2. Pelagic sediments (Boba Beds) consist predominantly of chert and calcilutite rich in radiolaria of Cretaceous age.

3. Batui Group consists of post-orogenic coarse clastics and volcanogenic turbidites of Late Miocene to Pliocene age.

4. Quaternary coralline reefs.

The oldest rock in the East Arm of Sulawesi is the Lemo Beds containing macroinvertebrate, including Misolia which typifies the Triassic sediments in eastern Indonesia.

The Jurassic succession is divided into 3 units, namely Lower Jurassic Kapali Beds, Late Jurassic Sinsidik Beds and Late Jurassic Nambo Beds containing *Belemnopsis uhligi* Stevens.

The Cretaceous sediments consist of two distinctive units, namely, 1. Boba Beds comprising abundant well-bedded chert and calcilutite rich in radiolaria of Valanginian (Early Cretaceous) to Late Cretaceous age. These rocks represent the pelagic cover of the ophiolite suite.

2. Luok Beds consisting predominantly calcilutite with chert nodules rich in microfossils of Late Cretaceous age.

The Early Tertiary sediments are dominated by carbonate platform, Salodik Limestones rich in benthic and planktonic foraminifera of Eocene to Miocene age.

The Neogene sediments consist of post-orogenic coarse clastics, the Batui Group, which includes the Kolo Beds, Biak Conglomerates and Lonsuit Turbidites. These rocks contain foraminifera of Late Miocene to Pliocene age.

B. Sedimentology, Petrology and Provenance

The vertical profile of sedimentary successions in the East Arm of Sulawesi can be divided into three different provenance and basinal settings:

1. Continental block derivation of detrital grains and lithic fragments is found within the Lemo Beds, Kapali Beds, Sinsidik Beds, Nambo Beds and Salodik Limestones.

Sedimentology of the Balantak Group suggests the sediments were deposited on the continental margin of the Banggai-Sula Platform.

The Triassic Lemo Beds represent channelised carbonate slope deposits. The Early Jurassic Kapali Beds were deposited in high energy shallow shelf. The Late Jurassic Nambo Beds and Sinsidik Beds were deposited in a neritic depth. The Late Cretaceous Luok Beds represent calcareous pelagic sediments deposited in subsided continental margin.

The Palaeogene Salodik Limestones are typical carbonate platform deposits rich in foraminifera.

2. Deep sea sediments dominated by radiolarian chert and calcilutite (Boba Beds) occur in association with and as part of the ophiolite suite. The occurrence of syngenetic manganese within the chert indicate the these rocks were deposited on oceanic crust beyond the reach of terrigenous clastic sedimentation.

3. Coarse clastic rocks (Batui Group) are typical post-orogenic sediments deposited on top of the collision complex. Detrital grains and rock fragments in Kolo Beds and Biak Conglomerates are wholly derived from collision complex, while those in the Lonsuit Turbidites point to a volcanic terrain derivation.

Sedimentology of the Batui Group indicates that these rocks were deposited in submarine fan depositional setting. The Kolo beds were deposited in outer fan and the Biak Conglomerates in a feeder canyon (channel) on the upper slope of trench-like basin. The Lonsuit Turbidites were deposited in an outer fan on forearc or arc trench gap basin. Megacyclic turbidites occur in the Lonsuit Turbidites. All these sediments show a very high ratio of sand to shale and a high proximality index (Pl) aś well.

C. Structures

The structural style and framework of the East Arm of Sulawesi are typical of a collision complex, resulting by collision of the northwest-moving Banggai-Sula Platform (BSP) and the eastern Sulawesi Ophiolite BElt (ESOB) in Middle Miocene time. The East Arm of Sulawesi collision complex consists of two distinctive structural domains :

1. Imbricated complex with an allochthonous Triassic to Palaeogene continental margin sequence (Balantak Group) juxtaposed with the ophiolite belt, and
2. Autochthonous Neogene coarse clastics and volcanogenic turbidites (Batui Group) and Quaternary coralline reefs.

The allochthonous rocks occur in fault-sliver or fault bounded exposures. Folding, fault and thrust of all scales occurred repeatedly in the imbricated complex.

The development of Balantak Duplex and Nambo Duplex is essentially related to the forward migration of the Batui Thrust during the later stage of overthrust of the ESOB onto the continental margin of the BSP.

The most prominent structures occurring in the East Arm of Sulawesi are the Batui Thrust and the Balantak Fault System, which are considered to be the surface expression of the collision zone between the BSP and ESOB. Kolokolo Melange contains fragments detached from the ophiolitic and continental margin sediments and matrix of calcareous mudstone with planktonic foraminifera of Middle Miocene to Pliocene age. Oil seeps also occur mostly within the fault zone.

The Balantak fault System is a dextral strike-slip fault produced by pure shear due to the NW-SE compression resulting by the northwest movement of the BSP. The development of left lateral Toili Fault Zone is also related to this pattern.

D. <u>Tectonics</u>

In summarising the tectonic history of the East Arm of Sulawesi three major tectonic events occurred during Early Mesozoic to Neogene times, which can be adequately explained in the light of plate tectonic theory.

1. Tectonic divergence of northern margin of the Australian continent in Triassic to Early Jurassic gives rise to the detachment and displacement of the BSP and the other microcontinents (EICF). Two extreme models for detachment of the EICF are possible : transcurrent displacement and rift-drift processes.

2. Tectonic convergence in Late Cretaceous - Paleocene is marked by the subduction of oceanic crust beneath the continental margin of Sundaland (Eurasia Plate). The formation of the CSMB, the development of the Palaeeogene Volcanic Arc in WSVB and the original displacement of the ESOB seem to have occurred during this tectonic convergence.

3. Tectonic convergence in Middle Miocene is dominated by collision of the ESOB and BSP. The emplacement of the Balantak Ophiolite and the development of the present configuration of the Sulawesi has occurred in this event. A double subduction zone might have been developed in northern Sulawesi : i.e. the southward dipping subduction zone in North Sulawesi Trench and the NNW-dipping subduc tion zone in the East Arm of Sulawesi.

5.8 RECOMMENDATIONS

In attempt to solve the regional problems, it is recommended for future work in the East Arm of Sulawesi and the adjacent areas :

1. In addition to the radiometric dating of the ophiolitic rocks (Balantak Ophiolite) presented in this thesis, it needs more systematic radiometric analysis of the ophiolitic rocks in the ESOB. The result will provide definite age for the ESOB and determine whether it is the result of the accretion of an extensive area of oceanic crust of a wide range of ages or a segment of oceanic crust of more restricted age with scattered seamounts formed at a later time.

2. Intensive field and biostratigraphic study of the pelagic cover of the ophiolite suite (chert and

calcilutite), systematically from the East Arm, Central Sulawesi, Southeast Arm, Kabaena and Buton Islands. The result will provide stratigraphic control especially of the stratigraphical relationship between the chert and calcilutite, and its age relationship to the ophiolite.

3. Systematic study (geochemistry and radiometric analysis) of volcanic arcs in the Western Sulawesi Volcano Plutonic Belt. The result will provide information about the volcanic environments, age and distribution of the Palaeogene and Neogene Volcanic Arcs.

4. Detailed study of metamorphic rocks in the region particularly those in the Central Sulawesi Metamorphic Belt especially age determinations and environments of formations.

5. Paleomagnetic analysis of the ophiolite and chert systematically from north to south of the ESOB.

6. Seismic Reflection studies would provide subsurface control of the collision complex in eastern Sulawesi, thickness of the ophiolite belt and the thickness and structure of the underplated continental margin sequence.

7. Gravity study across the East Arm and North Arm of Sulawesi, across Central Sulawesi, and across Buton, Southeast Arm, Bone Gulf and South Arm of Sulawesi would provide control on crustal modelling across the collision complex.

REFERENCES

Aalto, K.R., 1982. The Franciscan Complex of northernmost California : Sedimentation and Tectonics. In : Leggett, J.K. (ed.) Trench-Forearc Geology, Geol. Soc. London Spec. Pub. **10**, 419-432.

Ahmad, W., 1975. Geology along the Matano Fault Zone, East Sulawesi, Indonesia. Proc. Reg. Conf. on the Geol. and Miner. Resour. of SE Asia, Jakarta, 143-150.

Ahr, W.M., 1973. The carbonate ramp; an alternative to the shelf model. Trans. Gulf. Cst. Ass. geo. Socs. **23**, 221-225.

Allemann, F. & Peters, T., 1972. The ophiolite-radiolarite belt of the North Oman Mountains. Eclogae Geol. Helvetiae **65**, 657-697.

American Geological Institute, 1960. Dictionary of Geological Terms, New York : Dolphin Book, 545 pp.

Arthur, M.A. & Premoli Silva, I., 1982. Development of widespread organic carbon-rich strata in the Mediterranean Tethys. In : Schlanger, S.O. & Cita, M.B. (eds.) Nature of Cretaceous carbon-rich facies. Academic Press, London, 7-54.

Audley-Charles, M.G., 1977. Mesozoic evolution of the margins of Tethys in Indonesia and Philippines: Indon. Petrol. Assoc. 5th Annual Convention, Jakarta, 1976. Proceedings, Vol. 2, 25-52.

----- 1978. The Indonesian and Philippines Archipelagoes. In: Moullade, M. & Nairn, A.E.M. (eds.) Phanerozoic Geology of the World. II. The Mesozoic,A. New York; Elsevier, 165-207.

----- 1981. Geological history of the region of Wallace's line. In: Whitmore, T.C. (ed.) Wallace's line and Plate Tectonics. Clarendon Press, Oxford, p. 25-35.

----- 1983. Reconstruction of eastern Gondwanaland. Nature, **306**, 48-50.

----- 1984. Cold Gondwana, warm Tethys and the Tibetan Lhasa Block. Nature, **310**, 165-166.

----- 1986. Rates of Neogene and Quaternary tectonic movements in the southern Banda Arc based on micropaleontology. J. Geol. Soc. London **143**, 161-175.

----- **& Milsom, J.S.,** 1974. Comment on 'Plate convergence, transcurrent faults and internal deformation adjacent to Southeast Asia and Western Pacific' by T.J. Fitch. J. Geophys. Res., **79**, 4980-81.

----- **Carter, D.J. & Milsom, J.S.,** 1975. Tectonic development of eastern Indonesia in relation to Gondwanaland dispersal. Nat. Phys. Sci. Jour., **239**, 35-39.

----- ------ **Barber, A.J., Norvick M.S. and Tjokrosapoetro, S.,** 1981a. Reinterpretation of the geology of Seram: Implication for the Banda Arcs and Northern Australia. In Barber, A.J. & Wiryosujono S. (eds.) The Geology and tectonics of eastern Indonesia. Geol. Res. & Develop. Centre, Bandung , Spec. Publ. **2**, 217-238.

Azwan, 1982. Geologic map of Bacan Quadrangle, Maluku, 1:250.000 scale. Open file report. Geol. Res. Dev. Centre, Bandung.

Bailes, A.H., 1980. Origin of Early Proterozoic volcaniclastic turbidites, south margin of the Kissenew sedimentary gneiss belt, File Lake, Manitoba. Precambrium Res. **12**, 197-225.

Bally, A.W., Watts, A.B., Grow, J.A., Manspeizer, W., Bernoulli, D., Schreiber, C. & Hunt, J.M., 1981. Geology of Passive Continental Margins; History, structure and sedimentologic Record. (with special emphasis on the Atlantic Margin). AAPG, Education Course Note Series 19. For the AAPG Eastern Section Meeting and Atlantic Margin Energy conference.

Barber A.J., 1979. Structural interpretations of the island of Timor. Proc. Southeast Asia Petrol. Expl. Soc. **4**, 9-21.

----- **Audley-Charles, M.G.,** 1976. The significance of the metamorphic rocks of Timor in the development of the Banda Arc, Eastern Indonesia. Tectonophysics **30**, 119-128.

----- ----- **Carter, D.J.,** 1977. Thrust tectonics in Timor. J. geol. Soc. Australia **24**, 51-62.

----- **& Wiryosujono, S., (eds.)** 1981. The Geology and tectonics of Eastern Indonesia. Geological Research and Development Centre, Bandung, Spec. Publ. 2, 217-238.

----- **Tjokrosapoetro, S. & Charlton, T.R.** (in press). Mud volcanoes, shale diapirs, wrench faults and melangess in accretionary complexes in eastern Indonesia. **AAPG.**

Barnes, R.P. & Andrews, J.R., 1984. Hot or cold emplacement of the Lizard Complex ? J. geol. Soc. London, **141**, 37-39.

Bemmelen, R.W. van, 1949. The Geology of Indonesia, Vol. IA, Government Printing Office, the Hague, 732 pp.

Ben-Avraham, Z., Nur, A. & Jones, D., 1982. The emplacement of ophiolites by collision. J. Geophys. Res. **87**, 3861-3867.

Berger, W.H. & Winterer, E.L., 1974. Plate stratigraphy and the fluctuating carbonate line. In : Hsu, K.J. & Jenkyns, H.C. (eds.) Pelagic sediments on land and under the sea. Spec. Pub. Int. Ass. Sedimentol. 1, 1-48.

Berner, R.A., 1970. Sedimentary pyrite formation. Am. J. Sci. **208**, 1-23.

Bernouli, D. & Jenkyns, H.C., 1974. Alpine, Mediterranean and central Atlantic Mesozoic facies in relation to the early evolution of the Tethys. In: Dott, R.H. Jr. & Shaver, R.H. (eds.). Modern and Ancient Geosynclinal Sedimentation. Soc. Econ. Paleont. Miner. Spec. Publ. **19**, 129-160.

----- & Lemoine, M., 1980. Birth and early evolution of Tethys : the overall situation. In : Aubouin, J., Debelmas, J. & Latrelle, M. (eds.) Geologie des chaines Alpines issues de la Tethys: France, Bureau Recherches Geologiques et Minieres, Memoire, **115**, 168-179.

Bentor, Y.K., 1980. Phosphorites - the unsolved problems. In: Bemtor, Y.K. (ed.) Marine Phosphorites : Geochemistry, Occurrence and Genesis. Soc. Econ. Pal. Miner. Spec. Publ. **29**, 3-18.

Blatt, H., 1982. Sedimentary Petrology. Freeman and Company,San Fransisco, 564 pp.

---------, and Christie, J.M., 1963. Undulatory extinction in quartz of igneous and metamorphic rocks and its significance in provenance studies of sedimentary rocks. J. Sed. Petrol. 33, 559-579.

--------, Middleton, G. & Murray, R., 1980. Origin of Sedimentary Rocks. Prentice-Hall, Inc. New Yersey, 634 pp.

Blow, W.H.,1969. Late Miocene to Recent planktonic foraminiferal biostratigraphy. Proc. First Int. Conf. Plank. Microfoss. **1**: 199-421.

Bonini, W.P., Loomis, T.P. & Robertson, J.D., 1973. Gravity anomalies, ultramafic intrusions, and the tectonics of the region around the Strait of Gibraltar. J. Geophys. Res. **78**, 1372-1382.

Bouma, A.H., 1962. Sedimentology of some flysch deposits. Amsterdam : Elsevier Pub. Co., 168 p.

Bouma, A.H., Normark, W.R. & Barnes, N.E. (eds.) 1985. Submarine Fans and Related Turbidite Systems. New York : Springer-Verlag, 351 pp.

Bowin, C.O., Lu, R.S., Lee, C.S. & Schouten, H., 1978. Plate convergence and accretion in the Taiwan-Luzon region. Amer. Ass. Petrol. Geol. Bull. **62**, 1645-1672.

Bowin, C., Purdy, C.M., Johnstone, C., Shor, G.G., Lawver, L., Hartono, H.M.S. & Jezek, P., 1980. Arc-continent collision in Banda Sea region. Amer. Assoc. Petrol. Geol. Bull. **64**, 868-915.

Brookfield, M.E., 1976. The emplacement of giant ophiolite nappes. I Mesozoic-Cenozoic examples. Tectonophysics, **37**, 247-303.

Browning, P., 1984. Cryptic variation within the cumulate sequence of the Oman Ophiolite: magma chamber depth and petrological implications. In: Ophiolites and oceanic lithosphere Gass, I.G., Lippard, S.J. & Shelton, A.W. (eds.). Geol. Soc. London Spec. Publ. **13**, 71-82.

Brouwer, H.A., 1930. The major tectonic features of Celebes. Proc. Kon. Akad. v. Western, Amsterdam, 338-343.

----- 1934. Geologische onderzoekingen op het eiland Celebes. Ver. Geol. Mijnb. Gen. Ned. & Kol. Geol. Serie **10**, 39-171.

----- 1947. Geological exploration in the island of Celebes. Geological summary and petrology, Amsterdam. Unpubl. report GRDC, Bandung, p. 1-64.

Burchette, T.P. & Britton, S.R., 1985. Carbonate facies analyses in the exploration for hydrocarbons : a case-study from the Cretaceous of the Middle East. In : Brenchley, P.J. & Williams, B.P.J. (eds.). Sedimentology : Recent development and applied aspects. Geol. Soc. London, 311-338.

Burke, G.A. & Drake, C.L., 1974. Geologic significance of continental margins. In: Burke, C.A. & Drake, C.L. (eds.) The Geology of Continental margins. New York; Springer-Verlag, p. 3-10.

Byers, C.W., 1977. Biofacies pattern in euxinic basins: a general model. In: Cook, H.E. & Enos, P. (eds.) Deep carbonate environments. Soc. Econ. Pal. Miner. Spec. Publ. 25, 5-17.

Cann, J.R., 1968. Geological processes at mid-ocean ridge crests. Geophys. J. Roy. Astron. Soc. **15**, 331-342.

----- 1969. Spilites from the Carlsberg Ridge, Indian Ocean. J. Petrol. **10**, 1-19.

----- 1970. New model for the structure of the ocean crust. Nature, London **226**, 928-930.

Cardwell, R.K. & Isacks, B.L., 1978. Geometry of subducted lithosphere beneath the Banda Sea in eastern Indonesia from seismicity and fault-plane solutions. J. Geophys. Res. **87**, 2825-2838.

----- **Kappel, E.S., Lawrence, M.S. & Isacks, B.L.**, 1981. Plate convergence along the Indonesian arc (abs.). EOS 62, 404.

Carter, D.J., Audley-Charles, M.G. & Barber, A.J., 1976. Stratigraphical analysis of island arc-continental margin collision in eastern Indonesia. J. Geol. Soc. London **132**,179-198.

Casey, J.F. & Karson, J.A., 1981. Magma chamber profiles from the Bay of Island Ophiolite Complex: implications for crustal-level magma chambers at mid-ocean ridges. Nature, London, **292**, 295-301.

----- **Fox, P.J., Karson, J.A. & Rosentcrantz, E.**, 1981. Heterogeneous nature of oceanic crust and upper mantle: a perspective from the Bay of Islands ophiolite. In : Emiliani, C. (ed.) The Sea, Vol. Vii. The Oceanic Lithosphere. New York; John Wiley, pp. 305-338.

----- **& Dewey, J.F.**, 1984. Initation of subduction zones along transform and accreting plate boundaries, triple junction evolution, and forearc spreading centres - implications for ophiolite geology and obduction. In :Gass, I.G., Lippard, S.J. & Shelton, A.W. (eds.), Ophiolites and Oceanic Lithosphere, Geol. Soc. London Spec. Publ.13, 269-290.

Chichester, A.H. & Cady, W.M., 1972. Origin and emplacement of Alpine-type ultramafic rocks. Nature Phys. Sci. **240**, 27-31.

Christensen, N.I. & Salisbury, M.H., 1975. Structure and constitution of the lower oceanic crust. Rev. Geophys. Space Physics **13**, 57-86.

Church, W.R., 1972. Ophiolite: its definition, origin as oceanic crust, and mode of emplacement in orogenic belts, with special reference to the Appalachians. Dept. Energy, Mines REs. Canada Publ., **42**, 71-85.

----- & **Stevens, R.K.**, 1971. Early Paleozoic Ophiolite complexes of the Newfoundland Appalachians as mantle-oceanic crust sequences. J.Geophys. Res., **76**, 1460-1466.

----- **& Riccio, L.**, 1977. Fractionation trends in the Bay of Islands ophiolite of Newfoundland: polycyclic cumulate sequences in ophiolites and their classification. Can. J. Earth Sci., **14**, 1156-1165.

Coleman, R.G., 1971. Plate tectonic emplacement of upper mantle peridotites along continental edges. J. Geophys. Res., **76**, 1212-1222.

-----, 1977. Ophiolites. Ancient Oceanic Lithosphere ? New

York; Springer-Verlag, 229 pp.

Coleman, R.G., 1981. Tectonic setting of ophiolite obduction in Oman. J. Geophys. Res., **86**, 2497-2508.

----- 1984. Ophiolites and tectonic evolution of the Arabian Peninsula. In: Gass, I.G., Lippard, S.J. & Shelton, A.W. (eds.)Ophiolites and Oceanic Lithosphere. Geol. Soc. London Spec. Publ. 3, 359-366.

Commission for Stratigraphic Code of Indonesia, 1975. Stratigraphic Code of Indonesia. Assoc. Indon. Geologists, Bandung, 19 p.

Conybeare, C.E.B., 1979. Lithostratigraphic Analysis of Sedimentary Basins. Academic Press, New York, 555 pp.

Crostella,A., 1977. Geosyncline and plate tectonics in Banda Arcs, eastern Indonesia. Amer. Assoc. Petrol. geol. Bull., **61**, 2063-2081.

Davies, H.L., 1976. Papua New Guinea Ophiolite. 25th Internat. Geol. Congress, Sydney, Excursion Guide 52A, 13pp.

----- & Smith, A.L., 1971. Geology of eastern Papua. Bull. Geol. Soc. Amer. **82**, 8299-8312.

----- & Jaques, A.L., 1984. Emplacement of ophiolite in Papua New Guinea. In : Gass, I.G., Lippard, S.J. & Shelton, A.W. (eds.) Ophiolites and Oceanic Lithosphere, Geol. Soc. London Spec. Publ. 3, 341-350.

Davies, T.A., Weser, O.E., Luyendyk, B.P. & Kidd, R.B., 1975. Unconformities in the sediments of the Indian Ocean. Nature, **253**, 15-19.

Dewey, J.F., 1974. Continental margins and ophiolite obduction: Appalachian Caledonian System. In : Burke, C.A., and Drake, C.L. (eds.)The Geology of Continental Margins. New York; Springer, pp. 933-950.

-----, 1976. Ophiolite obduction. Tectonophysics 31, 93-120.

-----, and Bird, J.M., 1970. Mountain Belts and the new global tectonics. J. Geophys. Res.. **75**, 2625-2647.

-----,------, 1971. Origin and emplacement of the ophiolite suite : Appalachian ophiolites in Newfoundland. J. Geophys. Res. **76**, 3179-3206.

-----, and Burke, 1974. Hot spots and continental break-up, implications for collision orogeny. Geology 2, 57-60.

Dickinson, W.R., 1970. Interpreting detrital modes of greywacke and arkose. J.Sed. Petrl. **40**, 695-707.

-----, 1971a. Plate tectonic model for orogeny at continental margins. Nature, London, 232, 41-42.

-----, 1971b. Plate tectonic models of geosynclines. Earth Plant. Sci. Lett. 10, 165-174.

-----, 1974. Tectonics and sedimentation. Soc. Paleont. Miner. Spec. Pub. 22.

-----, **and Seely, D.R.,** 1979. Structure and stratigraphy of fore-arc regions. AAPG. Bull., 63, 2-31.

----- **and Suczek, C.A.** 1979b. Plate tectonics and sandstone compositions. AAPG. Bull. 63,2164-2182.

Dietz, R.S. & Holden, J.C., 1970. Reconstruction of Pangea: Break up and dispersion of continents, Permian to present. J. Geophys. Res., 75, 285-305.

Djuri, H.M. & Sudjatmiko, 1974. Geologic map of the Majene and western part of the Palopo Quadrangle, South Sulawesi, 1:250.000 scale. Geological Survey of Indonesia, Bandung.

Dott, R.H.Jr., 1964. Wacke, graywacke and matrix - What approach to immature sandstone classification. J. Sed. Petrol. 34, 625-632.

Dott , R.H. Jr., 1978. Tectonics and sedimentation a century later. Earth Sci. Rev. 14, 1-34.

Dott, R.H.Jr. & Shaver, R.H. (eds.), 1974. Modern and Ancient Geosynclinal Sedimentation. Soc. Econ. Paleont. Miner. Spec. Pub. 19.

Dow, D.B., 1977. A geological synthesis of Papua New Guinea. Bull. Bur. Miner. Resour. Geol. Geophys. Aust. 201, 1-41.

------, **and Sukamto, R.,** 1984. Western Irian Jaya : The end-product of oblique plate convergence in the Late Tertiary. Tectonophysics 106, 109-139.

Dunham, R.J., 1962. Classification of carbonate rocks according to depositional texture. In : Ham, W.E. (ed.) Classification of Carbonate Rocks, 108-121. Am. Assoc. Petrol. Geol. Oklahoma.

Eisbacher, G.H., 1974. Evolution of successor basins in the Canadian Cordillera. In : Dott, R.H.Jr and Shaver, R.H. (eds.) Modern and Ancient Geosynclinal Sedimentation, Soc. Econ. Paleon. Miner. Spec. Publ. 19, 274-291.

Elthon, D., Casey,J.F. & Komar, S., 1982. Mineral chemistry of ultramafic cumulates from the North Arm Mountain massif of the Bay of Islands Ophiolite: evidence for high-pressure crystal fractionation of oceanic basalts. J. Geophys. Res.

87, 8717-8734.

Elter, P. & Trevisan, L., 1973. Olistostromes in the tectonic evolution of the Northern Appennines. In : De Jong, K.A. & Scholten, R. (eds.). Gravity and Tectonics New York: Wiley, pp. 175-188.

England, R.N. & Davies, H.L., 1973. Mineralogy of ultramafic cumulate and tectonites from eastern Papua. Earth Planet. Sci. Lett. **17**, 416-425.

Evans, C.A., Hawkins, J.W. & Moore, G.F., 1982. Petrology and geochemistry of ophiolitic and associated volcanic rocks on the Talaud Islands, Molucca Sea collision zone, NE Indonesia. In : Hilde, T. (ed.) Geodyn. Ser. AGU., Washington DC.

Evans, I. & Kendall, C.G.St.C., 1977a. An Interpretation of depositional setting of some deep-water Jurassic carbonates of the Central High Atlas Mountains, Morocco. In : Cook, H.E. & Enos, P. (eds.) Deep-water Carbonate Environments, Soc. Econ. Paleon. Miner. Spec. Publ. 25, 249-261.

----- ----- **& Butler, J.C.**, 1977b. Genesis of Liassic shallow water rhythms, Central High Atlas Mountains, Morocco. J. Sed. Petrol. **47**, 120-128.

Falvey, D.A., 1974. The development of continental margins in plate tectonic theory. APEA Jour. **14**, 95-106.

----- **& Mutter, J.C.**, 1981. Regional plate tectonics and the evolution of Australian passive continental margins. Bur. Miner. Res. J. Aust. Geol. Geophys. 6, 1-29.

Fisher, R.V., 1979. Models for pyroclastic surges and pyroclastic flows. J. Volcan. geoth. Res. Amsterdam 6, 305-318.

-----, 1983. Flow transformation in sediment gravity flows. Geology **11**, 273-274.

----- 1984. A review of submarine volcanism, transport process and deposits. In : Kokelaar, B.P. & Howells, M.F. (eds.) Marginal Basin Geology: Volcanic and Associated Sedimentary and Tectonic Process in Modern and Ancient Marginal Basins, Geol. Soc. London Spec. Publ. 16, 15-28.

----- **& Schmincke, H.U.**, 1984. Pyroclastic Rocks. Heidelberg; Springer.

Fitch, T.J., 1970. Earthquake mechanisms and island-arc tectonics in Indonesia-Philippine region. Bull. Seismol. Soc. Amer. **60**, 565-591.

Folk, R.L., 1954. The distinction between grain size and mineral composition in Sedimentary Rocks Nomenclature. J, Geol. **62**, 344-459.

Folk, R.L., 1962. Spectral subdivision of limestone types. In : Ham, W.R. (ed.) Classification of carbonate rocks, pp. 62-84. Amer. Assoc. Petrol. Geol. Oklahoma.

Folk, R.L., 1968. Petrology of Sedimentary Rocks. Austin, Tex. : Hemphill's Book Store, 170 pp.

Fuchs, G., 1979. On the Geology of western Ladakh. Jahrab. Geol. B-A, 513-540.

Garrels, R.M. & Mackensie, F.T., 1971. Evolution of Sedimentary Rocks. New York; Norton.

Garrison, R.E., 1974. Radiolarian cherts, pelagic limestone and igneous rocks in eugeosynclinal settings. In : Hsu, K.J. & Jenkyns, H.J. (eds.) Pelagic Sediments: On Land and Under the Sea. Internat. Assoc. Sedimentol. Spec. Publ. 1, 367-399.

----- & Fisher, A.G., 1969. Deep-water limestones and radiolarites of the Alpine Jurassic. Soc. Econ. Paleont. Miner. Spec. Pub. **14**, 20-55.

Gass, I.G., 1963. Is the Troodos Massif of Cyprus a Mesozoic ocean floor ? Nature **220**, 39-42.

---- & Masson-Smith, D., 1963. The geology and gravity anomalies of the Troodos Massif, Cyprus. Roy. Soc. London Philos. Trans. Ser. **A255**, 417-467.

---- & Smewing, J.D., 1973. Intrusion, extrusion and metamorphism at constructive margins: evidence from the Troodos Massif, Cyprus. Nature, **242**, 26-29.

---- Lippard, S.J. & Shelton, A.W., 1984. Ophiolites and Oceanic Lithosphere. Geol. Soc. London Spec. Publ. 13, 413 pp.

Gealey, W.K., 1977. Ophiolite obduction and geologic evolution of the Oman Mountains and adjacent areas. Bull. geol. Soc. Am. **88**, 1183-1191.

Gibbs, A.D., 1984. Structural evolution of extensional basin margins. J. Geol. Soc. London, Vol. **141**, 609-620.

Greensbaum, D., 1972. Magmatic processes at ocean ridges; evidence from the Troodos Massif, Cyprus. Nature, Phys. Sci. **238**, 18-21.

Gregory, R.T., 1984. Melt percolation beneath a spreading ridge : evidence from the Semail peridotites, Oman. In :

Gass, I.G., Lippard, S.J. & Shelton, A.W. (eds.) Ophiolites and Oceanic Lithosphere. Geol. Soc. London Spec. Pub. 13, 55-62.

Gribi Jr, E.A., 1973. Tectonics and Oil prospects of the Moluccas, Eastern Indonesia. Geol. Soc. Malay. Bull., 6, 11-16.

Haile, N.S., 1978. Paleomagnetic evidence for the rotation of Seram, Indonesia. J. Geophys. of the Earth, vol. **26** supplement., S191-S198.

Hall, R., 1984. Ophiolites : figments of oceanic lithosphere ? In : Gass, I.G., Lippard, S.J. & Shelton, A.W. (eds.) Ophiolites and Oceanic lithosphere. Geol. Soc. London Spec. Pub. 13, 393-404.

Hallam, A., 1975. Jurassic Environments. Cambridge University Press, 269 pp.

----- & **Bradshaw, M.J.**, 1979. Bituminous shales and oolitic ironstones as indicator of transgression and regression. J. geol. Soc. London **136**, 157-164.

Hamilton, W.H., 1973. Tectonics of the Indonesian Region. Geol. Soc. Malay. Bull. **6**, 3-10.

----- 1977. Subduction in the Indonesia region. In: Island Arcs, Deepsea Trenches and Back Arc Basins, Maurice Ewing Series. Amer. Geophys. Union 1, 15-31.

----- 1978. Tectonic map of the Indonesia region. U.S. Geol. Survey Map I-875-D. Miscellaneous Investigation Series, 1 sheet 1:5,000,000.

-----, 1979. Tectonics of the Indonesian Region. US. Geol. Surv. Prof. Paper., 1078, Washington, 345 pp.

Hampton, M.A., 1972. The role of subaqueous debris flow in generating turbidity currents. J. Sed. Petrol. **42**, 775-793.

----- 1975. Competence of fine-grained debris flows. J. Sed. Petrol. **45**, 834-844.

Harland, W.B., Cox, A.V., Llewellyin, P.G., Pickton, C.A.G., Smith, A.G. & Walters, R., 1982. A Geologic Time Scale. Cambridge Univ. Press, 131 pp.

Hashimoto, W., Aliate, E., Aoki, N., Balee, G., Ishibasi, T., Kitamura, N., Matsumato, T., Tamura, M. & Yanagida, J., 1975. Cretaceous system of southeast Asia. Geol. Paleont. SE.Asia, 15, 219-280.

Hatch, F.H., Rastall, R.H. & Greensmith, J.T., 1971.

Petrology of Sedimentary Rocks. London : Murby.

Hatherton, T. & Dickinson, W.R., 1969. The relationship between andesitic volcanism and seismicity in Indonesia, the Lesser Antilles and other island arcs. J. Geophys. Res. **74**, 5301-5310

Haynes, J.R., 1981. Foraminifera. London : MacMillan Pub.Ltd., 433 pp.

Hedberg, H.D. (ed.), 1976. International Stratigraphic guide. A Guide to stratigraphic classification, terminology, and procedures. International Union of Geological Sciences, Commission on Stratigraphy, International Subcommission on Stratigraphic classification. New York : Wiley, 200 pp.

Heezen, B.C., Tharp, M. & Ewing, M., 1959. The floor of the oceans. 1. The North Atlantic. Geol. Soc. Am. Spec. Paper **65**, 122p.

Heirtzler, J.R., Cameron, P., Cook, P.J., Powell, T., Roeser, H.A. Sukaradi, S. & Veevers, J.J., 1978. The Argo Abyssal Plain. Earth Plant. Sci. Lett. **41**, 21-31.

Hermes, J.J., 1968. The Papuan geosyncline and the concept of geosynclines. Geol. Mijnb. **47**, 81-97.

Hilde, T.W.C., Uyeda, S. & Kroenke, L., 1977. Evolution of the western Pacific and its margin. Tectonophysics **38**, 145-165.

Hiscott, R.N. & Middleton, G.V., 1979. Depositional mechanics of thick bedded sandstones at the base of a submarine slope, Tourella Formation (Lower Ordovician, Quebec, Canada. Spec. Publ. Soc. Econ. Paleont. Miner. Tulsa **27**, 307-326.

Hobbs, B.E., Means, D.W. & Williams, P.F., 1976. An Outline of Structural Geology. New York : John Wiley & Sons, 571 pp.

Holcombe, C.J., 1977. How rigid are the lithospheric plates ? Fault and shear rotation in Southeast Asia. J. Geol. Soc. London **134**, 325-342.

Hopper, R.H., 1941. A geological reconnaisance in the east arm of Celebes and on the island of Peleng. Unpubl. Report NPPM.

Hopson, C.A., Coleman, R.G., Gregory, R.T., Pallister, J.S. & Bailey, E.H., 1981. Geologic section through the Semail ophiolite and associated rocks along a Muscat-Ibra transect, Southeastern Oman Mountains. J. Geophys. Res. **86**, 2527-2544.

-----, 1968. Principles of melanges and their bearing on the Franciscan-Knoxville Paradox. Geol. Soc. Am. Bull. **79**, 1063-1074.

Hsu, K.J., 1974. Melanges and their distinction from olistostromes. In : Dott, R.H.Jr. & Shaver, R.H. (eds.) Modern and Ancient Geosynclinal Sedimentation. SEPM. Spec.Pub. **19**, 321-333.

----, and Bernoulli, D., 1978. Genesis of the Tethys and the Mediterranean : In : Hsu, K.J., Montadert, L. and others (eds.) Initial Reports of the Deep Sea Drilling Project **42**, US Govt. Printing Office, Washington, p. 943-949.

Hutchison, Ch.S., 1975. Ophiolite in Southeast Asia. Geol. Soc. Am. Bull. **86**, 797-806.

Irvine, T.N. & Findlay, T.C., 1972. Alpine-type peridotite with particular reference to the Bay of Islands complex. Pub. Earth Physics Branch Dept. Energy, Mines Res. Can. **42**, 97-128.

ISST: International Subcommission on Stratigraphic Terminology, 1965. Definition of geologic systems. Am. Ass. Petro. Geol. Bull. **49**, 1694-1703.

Jacobson, R.S., Shor, G.G.Jr., Kieckhefer, R.M. & Purdy, G.M., 1978. Seismic refraction and reflection studies in the Timor-Aru Trough system and Australian continental shelf. Am. Ass. Petrol. Geol. Memoir **29**, 209-222.

Jacques, A.L. & Robinson, G.P., 1977. The continent/island arc collision in northern Papua New Guinea. BMR. J. Austr. Geol. Geophys. **2**, 289-303.

James, N.P., 1979. Facies Models : Reefs. In : Walker, R.G. (ed.) Facies Models. Geol. Assoc. Canad. Geosci. Ser. **1**, 121-132.

Jenkyns, H.C., 1980a. Tethys : past and present. Proceedings of the Geol. Assoc. **91**, 107-118.

------, 1980b. Cretaceous anoxic events : from continents to oceans. J. geol. Soc. London **137**, 171-189.

Johnson, B.D., Powell, C.McA. & Veevers, J.J., 1976. Spreading history of the eastern Indian ocean and Greater India's northward flight from Antarctica and Australia. Geol. Soc. Am. Bull. **87**, 1560-1566.

Johnston, C.R., 1981. A review of Timor tectonics with implications for the development of the Banda Arc. In : Barber, A.J. & Wiryosujono, S. (eds.) The Geology and Tectonics of Eastern Indonesia. Geol. Res. Dev. Centre Spec. Pub. **2**, 199-216.

Karig, D.E., 1971. Origin and development of marginal basins in the western Pacific. J. Geophys. Res. 76, 2542-2561.

-----, 1973. PLate convergence between the Philippines and Ryuku Islands. Marine Geology **14**, 153-168.

Karson, J. & Dewey, J.F., 1978. Coastal complex, western Newfoundland, an Early Ordovician fracture zone. Bull. geol. Soc. Am. **89**, 1037-1049.

Kastner, M. & Stonecipher, S.A., 1974. Zeolites in pelagic sediments of the Atlantic, Pacific and Indian Oceans. In : Sand, L.B. & Hampton, F.A. (eds.) Natural Zeolites : Occurrence, Properties and Uses, Oxford : Pergamon Press, p. 199-220.

Katili, J.A., 1970. Large transcurrent faults in Southeast Asia with special reference to Indonesia. Geol. Rundschau, **59**, 581-600.

-----, 1971. A review of geotectonic theories and tectonic maps of Indonesia. Earth Sci. Reviews **7**, 143-163.

-----, 1972. On fitting certain geological and geophysical features of the Indonesian island arcs to the new global tectonics. Univ. Western Austr. Press, p. 287-305.

-----, 1974. Geological environment of the Tndonesian mineral deposits - a plate tectonic approach. Pubk. Tekn. Ser. Geol. Ekon., Geol. Surv. Indon. **7**, 1-18.

-----, 1975. Volcanism and plate tectonics in the Indonesian island arcs. Tectonophysics **26**, 165-188.

-----, 1978. Past and present geotectonic position of Sulawesi, Indonesia. Tectonophysics **45**, 289-322.

-----,, 1985. Tectonic framework, earth resources and regional geological programme. Internat. Geol. Sci. Pub. 13, 18-68.

----- & Hartono, H.M.S. ,1983. Complications of Cenozoic tectonic development in eastern Indonesia. In : Hilde, T.W.C. & Uyeda, S. (eds.) Geodynamics of the western Paific - Indonesia region. Geodynamic Ser. 11, 387-399.

-----& Asikin, S., 1985. Hydrocarbon prospects in complex Paleo Subduction Zones. Proceed. Indon. Petrol. Assoc., 83-103.

Klompe, Th.H.F., 1956. The structural importance of the Sula Spur. Proc. 8th Pacific Sci. Cong. J. IIA, 869-888.

Koolhoven, W.C.B., 1930. The geology of the Malili area, Central Celebes. Jaar. Mijn. III, 127-153.

Koswara, A. & Sukarna, D., 1986 (in prepar.) Geologic map of Tukang Besi Island, Southeast Sulawesi, 1:250.000 scale. Geol. Res. Dev. Centre, Bandung.

Krause, D.C., Tectonics, marine geology and bathymetry of the Celebes Sea-Sula Sea region. Geol. Soc. Am. Bull. **77**, 813-832.

Krumbein, W.C. & Sloss, L.L., 1963. Stratigraphy and Sedimentation. San Francisco : Freeman.

Kundig,E., 1956. Geology and ophiolite problems of East Sulawesi. Verh. Kon. Ned. Geol. Mijnb. Gen. Geol. Ser. Jour. **XVI**, 210-235.

Lajoie, J., 1979. Facies Models of Volcanic Rocks. In : Walker, R.G. (ed.) Facies Models, Geol. Ass. Canad. Geosci. Canad. Ser. 1, 191-200.

Lapouille, A. 1985. Age and origin of the seafloor of the Banda Sea (Eastern Indonesia), Oceanologica Acta, **8**,379-389.

Larson, R.L., 1975. Late Jurassic sea-floor spreading in the eastern Indian Ocean. Geology 3, 69-71.

-----, 1977. Early Cretaceous breakup of Gondwanaland off Western Australia. Geology **5**, 57-60.

Laubscher, H. & Bernoulli, D., 1977. Mediterranean and Tethys. In : Nairn, A.E.M., Kanes, W.H. & Stehli, F.G. (eds.) The Ocean Basins and Margins. Plenum Publ. Corp. **4A**, 1-28.

Leeder, M.R., 1982. Sedimentology : Process and Product. London : George Allen & Unwin, 344p.

Leggett, J.K., 1982 (ed.). Trench-Forearc Geology : Sedimentation and Tectonics in Modern and Ancient Active Plate Margins. Geol. Soc. London Spec. Pub. **10**

Leggett, J.K., 1985. Deep-sea pelagic sediments and palaeo-oceanography : a review of recent progress. In : Brenchley, P.J. & Williams, B.P.J. (eds.) Sedimentology : Recent development and Applied Aspects, Geol. Soc. London, 95-122.

Leuwen, Th.M. van, 1981. The geology of southwest Sulawesi with special reference to the Biru area. In : Barber, A.J. & Wiryrosujono, S. (eds.) The Geology and Tectonics of eastern Indonesia, Geol. Res. Dev. Centre Spec. Pub. 2, 277-304.

Lockwood, J.P., 1971. Sedimentary and gravity-slide emplacement of serpentinite. Geol. Soc. Am. Bull. 82, 919-936.

Loczy, L. von, 1934. Geologie van noord Boengkoe en het Bongkagebuis tuschen de Golf van Tomini en de Golf Tolo in oost Celebes. Verh. Geol. Mijnb. Genootsch. Ned. en Kolonien, Geol. Serie **10**, 219-224.

Lofting, M.J.W., Crostella, A. & Halse, J.W., 1975. Exploration results and future prospects in the northern Australian region. World Petroleum Congress, 9th Tokyo, Proceedings 3, 65-81.

Loomis, T.P., 1972. Tertiary mantle diapirism, orogeny, and plate tectonics east of the Strait of Gibraltar. Am. J. Sci. **275**, 1-30.

Lovell, J.P.B., 1970. The palaeogeographical significance of lateral variations in the ratio of sandstone to shale and other features of Aberystwyth Grits. Geo. Mag. **107**, 147-158.

Mackenzie, W.S. & Gailford, C., 1982. Atlas of Rock-Forming Minerals in Thin Section. London : Longman, 98 pp.

Magaritz, M. & Taylor, H.P.Jr., 1974. Oxygen and hydrogen isotope studies of serpentinization in the Troodos ophiolite complex, Cyprus. Earth Planet. Sci. Lett. **23**, 8-14.

Malpas, J., 1979. Dynamothermal aureole beneath the Bay of Islands ophiolite in western Newfoundland. Canad. J. Earth Sci. **16**, 2086-2101.

----- **& Stevens, R.K.**, 1977. The origin and emplacement of the ophiolite suite with examples from western Newfoundland. Geotectonics **11**, 453-466.

----- **& Langdom, G.**, 1984. Petrology of the upper pillow lava suite, Troodos ophiolite, Cyprus. In : Gass, I.G., Lippard, S.J. & Shelton, A.W. (eds.) Ophiolites and Oceanic Lithosphere, Geol. Soc. London Spec. Pub. **13**, 155-167.

Masters, B.A., 1977. Mesozoic planktonic foraminifera : a world-wide review and analysis. In : Ramsay, A.T.S. (ed.) Oceanic Micropaleontology **1**, 301-731. New York : Academic Press.

Maxwell, J.C., 1970. The Mediterranean, ophiolites and continental drift. In : Johnson, H.S. & Smith, B.L. (eds.) Megatectonics of continents and oceans. New Brunswick, New Jersey : Rutgers Univ., pp. 167-193.

-----, 1973. Ophiolites - old oceanic crust or internal diapirs ? In : Symp. Ophiolites in the Earth's crust, Moscow : Acad. Sci. USSR, pp. 71-73.

-----, 1974. Anatomy of an orogen. Geol. Soc. Am. Bull. 85, 1195-1204.

McBride, E.F., 1963. Classification of common sandstones. J. Sed. Petrol. 33, 664-669.

McCaffrey, R., 1981. Crustal structure and tectonics of the Molucca Sea collision zone, Indonesia. Unpub. Thesis, University of California, St. Cruz, 157 pp.

-----, 1982. Lithospheric deformation within the Molluca Sea arc-arc collision : evidence from shallow and intermediate earthquake activity. J. Geophys. Res., **87**, 3662-3678.

-----, Sutardjo, R., Susanto, R., Buyung, R., Sukarman, B., Husni, M., Sudiono, Satuju, D. & Sukamto, E., 1983. Microearthquake surveys of the Molluca Sea and Sulawesi, Indonesia. Geol. Res. Dev. Cent. Bull. 7, 13-23.

-----, Silver, E.A. & Raitt, R.W., 1980. Crustal structure of the Molluca Sea collision zone, Indonesia. In : Hayes, D.E. (ed.) The Tectonic and Geologic Evolution of SE Asia Seas and Islands. Am. Geophys. Uni. Geophys. Mon. 23, 161-177.

----- ----- -----, 1981. Seismic refraction in the East Arm, Sulawesi - Banggai Islands region of Eastern Indonesia. In : Barber, A.J. & Wiryosujono, S. (eds.) The Geology and Tectonics of eastern Indonesia. Geol. Res. dev. Cent. Spec. Pub. 2, 321-326.

McClay, K.R. & Price, N.J. (eds.), 1981. Thrust and Nappe Tectonics. Geol. Soc. London Spec. Pub. 9, 539 pp.

McIlreath, I.A. & James, N.P., 1979. Facies Models: carbonate slopes. In : Walker, R.G. (ed.) Facies Models. Geol. Ass. Canad. Geosci. Canad. Ser. 1, 133-144.

McRae, S.G., 1970. Glauconite. Earth Sci. Rev. 8, 397-440.

Menzies, M., 1974. Plagioclase lherzolite-residual mantle relationships within two eastern Mediterranean ophiolites. Contr. Miner. Petrol. **45**, 197-213.

Menzies, M. and Allen, C., 1974. Plagioclase lherzolite-residual mantle relationships within two eastern Mediterranean ophiolites. Contr. Miner. Petrol., **45**, 197-213.

Miall, A.D., 1977. A review of the braided-river depositional environment. Earth Sci. Rev. 13, 1-62.

-----, 1978. Tectonic setting molasse and other non-marine paralic sedimentary basins. Canad. J. Earth Sci. **15**, 1613-1632.

-----, 1986. Eustatic sea level changes interpreted from seismic stratigraphy : A critique of the methodology with particular reference to the North Sea Jurassic record. AAPG **70**, 131-137.

Milsom, J.S., 1973. Papua Ultramafic belt : gravity anomalies and the emplacement of ophiolites. Geol. Soc. Am. Bull. **84**, 2243-2258.

-----, 1981. Neogene thrust emplacement from a frontal arc in New Guinea. In : McClay, K.R. & Price, N.J. (eds.) Thrust and Nappe Tectonics. Geol. Soc. London Spec. Pub. 9, 417-426.

-----, 1984. The gravity field of Marum ophiolite complex, Papua New Guinea. In : Gass, I.G., Lippard, S.J. & Shelton, A.W. (eds.). Ophiolites and Oceanic Lithosphere. Geol. Soc. London Spec. Pub. 13, 351-357.

-----, Audley-Charles, M.G., Barber, A.J. & Carter, D.J., 1983. Geological-geophysical paradoxes of the eastern Indonesia collision zone. In : Hilde, T.W.C. & Uyeda, S. (eds.) Geodynamics of the western Pacific-Indonesian region, Geodyn. Ser. Geophys. Union **11**, 401-411.

Mitchell, A.A.G., 1970. Facies of an Early Miocene volcanic arc, Malekula Island, New Hebrides. Sedimentology **14**, 201-243.

Miyashiro, A., 1973. The Troodos ophiolitic complex was probably formed in an island arc. earth Planet. Sci. Lett. **19**, 218-224.

Molnar, P. & Tapponnier, P., 1975. Cenozoic tectonics of Asia-effects of a continental collision. Science **189**, 419-426.

Moody, J.D. & Hill, M.J., 1956. Wrench-fault tectonics. Geol. Soc. Am. Bull. **67**, 1207-1246.

Moore, D.G., Curray, J.R. & Emmel, F.J., 1976. Large submarine slide (olistostrome) associated with Sunda Arc subduction zone, northeast Indian Ocean. Marine Geology **21**, 211-226.

Moore, J.G., 1975. Mechanism of formation of pillow lava. Am. J. Sci. **63**, 269-299.

Moore, T.C.Jr., Van Andel, Tj.H., Sancetta, C. & Pisias, N., 1978. Cenozoic hiatuses in pelagic sediments.

Micropaleontology, 24, 113-138.

Moores, E.M., 1982. Origin and emplacement of ophiolites. Rev. Geophys. Space Phys. **20**, 735-760.

----- & Vine, F.J., 1971. Troodos Massif, Cyprus and other ophiolites as oceanic crust : evolution and implications. Roy. Soc. London Philos. Trans. **A268**,443-466.

Mutti, E., 1977. Distinctive thin-bedded turbidite facies and related depositional environments in the Eocene Hecho Group (South-central Pyrenees, Spain). Sedimentology **24**, 107-131.

Mutti, E., Lucci, F.R., Seguret, M. & Zanzucchi, G., 1984. Seismoturbidites : A new group of resedimented deposits. Marine Geology **55**, 103-116.

Neef, G., Plimer, I.R. & Bottrill, R.S., 1985. Submarine-fan deposited sandstone and rudite in a Mid-Cenozoic interarc basin in Maewo, Vanuatu (New Hebrides)

Nilsen, T.H., Walker, R.G. & Normark, W.R., 1980. Modern and ancient submarine fans : discussion and replies. AAPG. Bull. **64**, 1094-1113.

Normark, W.R. & Piper, D.J.W., 1972. Sediments and growth patterns of Navy deep sea fan, San Clemete Basin, California Borderland. J. Geol. **80**, 198-223.

Norvick, M.S., 1979. The tectonic history of the Banda Arcs, eastern Indonesia : a review. J. Geol. Soc. **136**, 519-527.

Nur, A. & Ben-Avraham, Z., 1977. Lost Pacifica continent. Nature **270**, 41-43.

-----, -----, 1983. Break-up and accretion tectonics. In : Hashimoto, M. & Uyeda, S. (eds.) Accretion Tectonics in the Circum-Pacific regions, Advanc. in Earth Planet. Sci., 3-20.

Otofuji, Y., Sasajima, S., Nishimura, S., Dharma, A. & Hehuwat, F., 1981. Paleomagnetic evidence for clockwise rotation of the Northern Arm of Sulawesi, Indonesia. Earth Planet. Sci. Lett. **54**, 272-280.

Owen, H.G., 1983. Atlas of continental displacement : 200 million years to present. A test of conventional and expanding earth models. Cambridge Univ. Press.

Oxburgh, E.R., 1972. Plate tectonics and continental collision. Nature, London **239**, 202-204.

Pallister, J.S., 1981. Structure of the sheeted dike complex of the Samail Ophiolite near Ibra, Oman. J. Geophys. Res. **86**, 2661-2672.

-----, 1984. Parent magmas of the Samail Ophiolite, Oman. In : Gass, I.G., Lippard, S.J. & Shelton, A.W. (eds.) Ophiolites and Oceanic Lithosphere. Geol. Soc. London Spec.Pub. 13, 63-70.

Pessagno, E.A.Jr., 1976. Radiolarian zonation and stratigraphy of the Upper Cretaceous portion of the Great Valley Sequence, California Coast ranges. Micropaleont. Spec. Pub. **2**, 1-96.

-----, 1977. Lower Cretaceous radiolarian biostratigraphy of the Great Valley Sequence and Franciscan Complex, California Coast Ranges. Cushman Foud. Foram. Res. Spec. Pub. **15**, 170 pp.

Pettijohn, F.J., 1975. Sedimentary Rocks. New York : Harper and Row.

-----, Potter, F.J. & Siever, R., 1972. Sand and Sandstones. New York : Springer-Verlag.

P.T. INCO Indonesia, 1972. Laterite deposits in Southeast Arm of Sulawesi. Unpub. Report Reg. Conf. geol. SE. Asia, Kuala Lumpur.

Pieters, P., Pigram, C.J., Trail, D.S., Dow, D.B., Ratman, N. & Sukamto, R., 1983. The stratigraphy of western Irian Jaya. Geol. Res. Dev. Cent. Bull. **8**, 14-48.

Pigram, C.J., 1986. Discussion : Western Irian Jaya the end-product of oblique plate convergence in the Late Tertiary. Tectonophysics **121**, 345-350.

-----, and Panggabean, H., 1983. Age of the Banda Sea, eastern Indonesia. Nature **301**, 231-234.

-----, -----, 1984a. Rifting of the northern margin of the Australian continent and the origin of some mircrocontinents in eastern Indonesia. Tectonophysics **107**, 331-353.

Platt, J.P. & Leggett, J.K., 1986. Stratal extension in Thrust Footwalls, Makram Accretionary Prism : implications for Thrust Tectonics. AAPG. Bull. **70**, 191-203.

Postuma, J.A., 1971. Manual of Planktonic Foraminifera. Amsterdam : Elsevier.

Potter, P.E., 1978. Petrology and chemistry of modern big river sands. J. Geol. **86**, 423-449.

Potter, P.E. & Pettijohn, E.J., 1977. Paleocurrents and Basin Analysis. Berlin : Springer-Verlag.

Powell, D.E., 1976. The geological evolution and of the continental margins off north-west Australia. APEA. J. **16**, 12-23.

Powell, C.McA. & Johnson, B.D., 1980. Constraints on the Cenozoic position of Sundaland. Tectonophysics 63, 91-109.

Price, N.J., 1966. Fault and Joint Development in Brittle and Semi-Brittle rocks. Oxford : Pergamon, 176 pp.

-----, and Audley-Charles, M.G., 1983. Plate rupture by hydraulic fracture resulting in over-thrusting. Nature, London **3006**, 572-575.

Quick, J., 1981. Petrology and petrography of the Trinity peridotite, an upper mantle diapir in the eastern Klamath Mountains, Northern California. J. Geophys. Res. **86**, 837-863.

Quilty, P.G., 1975. Late Jurassic to Recent geology of the western margin of Australia. In : Veevers, J.J. (ed.) Deep Sea Drilling in Australia Waters, Challenger Symp., p. 15-23.

-----, Cenozoic sedimentation cyclic in western Australia. Geology **5**, 336-340.

-----, 1980. Sedimentation cycles in the Cretaceous and Cenozoic of western Australia. Tectonophysics **63**, 349-366.

Ramsay, J.G., 1967. Folding and Fracturing of Rocks. New York : McGraw-Hill.

Ratman, N., 1986. (in prep.) Geologic map of Mamuju Quadrangle, South Sulawesi, scale 1:250.000. Geol. Res. Dev. Centre, Bandung.

Reading, H.G., 1972. Global tectonics and the genesis of flysch successions. 24th Int. Geol. Congr. Proc. Sect. **6**, 59-66.

-----, 1978. Sedimentary Environments and Facies. London : Blackwell Sci. Pub., 569 pp.

-----, 1980. 'Characteristics and recognition of strike-slip fault system'. In : Balance, P.E. & Reading, H.G. (eds.) Sedimentation in oblique-slip mobile zone. Int. Assoc. Sediment. Spec. Pub. **4**, 7-26.

Ricci-Lucci, F., 1975. Depositional cycles in two turbidite formations of the Northern Appennines. J. Sed. Petrol. **45**, 3-43.

Robertson, A.H.F. and Woodcock, N.H., 1980. Strike-slip related sedimentation in the Antalya Complex, SW Turkey.

Spec. Publ. Int. Assoc. Sedimentol. **4**, 137-145.

Robinson, P.T., Melsom, W.G., O'Hearn, T. & Schmincke, H.U., 1983. Volcanic glass composition of Troodos Ophiolite, Cyprus. Geology **11**, 400-404.

Rod, E., 1974. Geology of Eastern Papua : discussion. Geol. Soc. Am. Bull. **85**, 633-658.

Roeder, D., 1977. Philippines arc system - collision or flipped subduction zone ? Geology 5, 203-206.

Roever, W.P., 1947. Igneous and metamorphic rocks in eastern Central celebes. Geol. Expl. in the island of Celebes, 65-173.

Rusmana, E., 1986 (in prep.). Geologic map of Kendari Quadrangle, SE. Sulawesi, scale 1:250.000. Geol. Res. Dev. Centre, Bandung.

Rusmana, E. & Hartono, U., 1983. Geologic map of Misol Quadrangle, scale 1:250.000. (Open file report). Geol. Res. Dev. Centre, Bandung.

Rusmana, E., Koswara, A. & Simandjuntak, T.O., 1984. Geologic map of Luwuk Quadrangle, Central Sulawesi, scale 1:250.000. (Open file rept.) Geol.Res. Dev. Centre, Bandung.

Salisbury, M.H. & Christensen, N.I., 1978. The seismic velocity structure of a traverse through the Bay of Islands Ophiolite complex, Newfoundland, an exposure of oceanic crust and upper mantle. J. Geophys. Res. 83, 805-817.

Sato, T., Westermann, G.E.G., Skwarko, S.K. & Hasibuan, F., 1978. Jurassic biostratigraphy of the Sula Islands, Indonesia. Bull. Geo. Surv. Indon. **4**, 1-28.

Sarasin, F. and Sarasin, P., 1901. Entwurf einer geografische, geologischen beschreibung der Insel Celebes. Weisbaden.

Scholl, D.W., Christensen, M.N., Huene, R. Von and Marlow, M.S., 1970. Peru-Chile trench sediments and sea-floor spreading. Geol. Soc. Am. Bull. **81**, 1339=1360.

-----, Huene, R. Von and Ridlon, J.B., 1968. Spreading of the ocean floor : undeformed sediments in the Peru-Chile trench. Science **159**, 869-871.

-----, Von Huene, R., Vallier, T.L. & Howell, D.G., 1980. Sedimentary masses and concepts about tectonic processes at underthrust ocean margin. Geology **8**, 564-568.

Searle, M.P. & Stevens, R.K., 1984. Obduction processes in ancient, modern and future ophiolites. In : Gass, I.G., Lippard, S.J. & Shelton, A.W. (eds.) Ophiolites and Oceanic Lithosphere. Geol. Soc. London Spec. Pub. 13, 303-319.

Sengor, A.M.C., 1979. Mid-Mesozoic closure of Permo-Triassic Tethys and its implications. Nature **279,** 590-593.

Sigurdson, H., Sparks, R.S.J., Carey, S.N. & Huang, T.C., 1980. Volcanogenic sedimentation in the Lesser Antilles Arc. J. Geoloogy **88,** 523-540.

Sikumbang, N. & Sanyoto, P., 1983. Geologic map of Buton Quadrangle, SE. Sulawesi, scale 1:250.000. (Open file report) Geol. Res. Dev. Centre, Bandung.

Silver, E.A., 1981. A new tectonic map of eastern Indonesia. In : Barber, A.J. & Wiryosujono, S. (eds.) The Geology and Tectonics of Eastern Indonesia. Geol. Res. Dev. Cent. Spec. Pub. 2, 343-347.

-----, Joyodiwiryo, Y.S. & McCaffrey, R., 1978a. Gravity results and emplacement geometry of the Sulawesi ultramafic belt, Indonesia. Geology **5,** 527-531.

----- & Moore, J.C., 1978b. The Molluca Sea collision zone, Indonesia. J. Geophys. Res. **83,** 1681-1691.

Simandjuntak, T.O., 1977. Megacyclic Turbidites. 7th Ann. Meet. Assoc. Indon. Geol., Bandung.

-----, 1980. Wasuponda Melange. 8th Ann. Meet. Assoc. Indon. Geol. Jakarta.

-----, 1981a. Some sedimentological aspects of Mesozoic rocks in eastern Sulawesi, Indonesia. 9th Ann. Meet. Assoc. Indon. Geol. Jogyakarta.

-----, 1981b. Sediment gravity flow deposits in Pangandaran-Cilacap region, southwest Java, and their bearing on the tectonic development of southwestern Indonesia. Geol. Res. Dev. Centre Spec. Pub. 2, 21-54.

-----, Rusmana, E. & Surono, 1981c (in press). Geologic map of Malili Quadrangle, South Sulawesi, scale 1:250.000. Geol. Res. Dev. Centre, Bandung

-----, ------, Supanjono, B.J., 1982 (in press). Geologic map of Bungku Quadrangle, Central Sulawesi, scale 1:250.000. Geol. Res. Dev. Centre, Bandung.

-----, Supanjono, B.J. & Surono, 1983 (in press). Geologic map of Poso Quadrangle, Central Sulawesi, scale 1:250.000. Geol. Res. Dev. Centre, Bandung.

-----, and Surono, 1983 (open file report). Geologic map of Kolaka Quadrangle, SE. Sulawesi, scale, 1:250.000. Geol. Res. Dev. Centre, Bandung.

-----, Christensen, N.I., Bartolomew, I.D. & Browning, P., 1984. The structure of the oceanic upper mantle and lower crust as deduced from the northern section of the Oman Ophiolite. In : Gass, I.G., Lippard, S.J. & Shelton, A.W. (eds.) Ophiolites and Oceanic Lithosphere, Geol. Soc. London Spec. Pub. **13**, 41-54.

Smith, A.G. & Woodcock, N.H., 1976. Emplacement model for some 'Tethyan' ophiolites. Geology **4**, 653-656.

-----, Hurley, A.M. & Briden, J.C., 1981. Phanerozoic Palaeo-continental World Maps. Cambridge Univ. Press.

Smith, C.H., 1958. Bay of Islands igneous complex, western Newfoundland. Geol. Surv. Canad. Mem. **290**, 1-32.

Spohn, T., 1981. On the origin of Orogenic Volcanism. Geol. Rundschau. **70**, 155-165.

Spoòner, E.T.C., 1974. Sub-sea-floor metamorphism, heat and mass tranfer; an additional comment. Contr. Miner. Petrol. **45**, 169-173.

Spray, J.G., 1984. Possible causes of upper mantle decoupling and ophiolite displacement. In : Gass, I.G., Lippard, S.J. & Shelton, A.W. (eds.) Ophiolites and Oceanic Lithosphere, Geol. Soc. London Spec. Pub. **13**, 255-268.

Stainforth, R.M., Lamb, J.L., Luterbacher, H. et al., 1975. Cenozoic planktonic foraminiferal zonation and characteristics of index forms. Contr. Paleont. Univ. Kansas, Art. **62**, 13-425.

Stanley, D.J., 1980. The Saint-Antonin Conglomerate in the Maritime Alps : a model for coarse sedimentation on a submarine slope : Smithsonian Contr. Marine Sci. **5**, 25 pp.

----- & Taylor, P.T., 1977. Sediment transport down a seamount flank by a combined current and gravity process. Marine Geol. **23**, 77-88.

-----, Palmer, H.D. & Dill, R.F., 1978. Coarse sedient transport by mass flow and turbidity current processes and downslope transformation in Annot Sandstone Canyon-Fan Valley systems. In : Stanley, D.J. & Kelling, E. (eds.) Sedimentation in Submarine Canyons, Fans and Trenches. Hutchinson & Ross, Stroudsburg, Penn., p. 85-115.

Stoneley, R., 1975. On the origin of ophiolite complexes in the southern Tethys region. Tectonophysics **25**, 303-322.

Storey, B. & MacDonald, D.I.M., 1984. Processes of formation and filling of a Mesozoic back-arc basin on the island of South Georgia. In : Kokelaar, B.P. & Howells, M.F. (eds.) Marginal Basin Geology : Volcanic and Associated Sedimentary and Tectonic Processes in Modern and Ancient Marginal Basin. Geol. Soc. London Spec. Pub. **16**, 5-28.

Stow, D.A.V. & Bowen, A.J., 1978. Origin of lamination in deep sea, fine grained sediments. Nature **274**, 324-327.

----- and Piper, D.J.W. (eds.), 1984. Fine-grained Sediments : Deep-water processes and facies. Geol. Soc. London Spec. Pub. **15**, 659 pp.

Sukamto, R., 1975a. Geologic Map of Indonesia, Sheet VIII, Ujung Pandang, 1:1000.000. scale. Geol. Surv. Indonesia.

-----, 1975b. Perkembangan tektonik di Sulawesi dan daerah sekitarnya, suatu sintesis perkembangan berdasarkan tektonik lempeng. Unpublished Geol. Surv. Indonesia.

-----, 1975c. The structure of Sulawesi in the light of plate tectonics. Reg. Conf. Geol. Miner. Res. SE.Asia, IAGI, Jakarta, 4-7.

-----, 1975d. Geologi daerah kepulauan Banggai dan Sula. Geol. Indon. **2**, 23-28.

-----, 1982a. Geologic map of Danau Tempe Quadrangle, South Sulawesi, 1:250.000 scale. Geol. Res. Dev. Centre, Bandung.

-----, 1982b. Geologic map of Lompobatang Quadrangle, South Sulawesi, scale 1:250.000. Geol. Res. Dev. Centre, Bandung..

-----, and Simandjuntak, T.O., 1983. Tectonic relationship between Geologic Province of Western Sulawesi, Eastern Sulawesi and Banggai-Sula in the light of sedimentological aspects. Geol. Res. Dev. Centre Bull. **7**, 1-12.

Sunarya, Y. & Yudawinata, K., 1980. Penelitian stratigrafi dan studi orientasi geokimia endapan bijih tipe Kuroko di daerah Sangkaropi, Kecamatan Sesean, Tana Toraja, Sulawesi Selatan. 9th Ann. Meet. Assoc. Indon. Geol. Yogjakarta.

Supanjono, J.B. & Haryono, E., 1986 (in prep.) Geologic Map of Taliabu Quadrangle, Banggai-Sula Islands. Geol. Res. Dev. Cntre, Bandung.

Suria-Atmadja, R., Golightly, J.P. & Wahyu, B.N., 1972. Mafic and ultramafic rock association in the East Arm of Sulawesi. Reg. Conf. Geol. SE. Asia, Kuala Lumpur.

Surono, 1981. Sedimen molasa di lengan timur Sulawesi. 10th Ann. Meet. Assoc. Indon. Geol. Bandung.

-----, Situmorang, R.L. & Simandjuntak, T.O., 1984 (Open file report). Geologic map of Batui Quadrangle, Central Sulawesi, 1:250.000 scale. Geol. Res. Dev. Centre, Bandung.

----- & Sukarna, E., 1986 (in prepar.) Geologic map of Sanana Quadrangle, Banggai-Sula Islands, 1:250.000 scale. Geol. Res. Dev. Centre, Bandung.

Tasse, N., Lajoie, L. & Dimroth, E, E., 1978. The anatomy and interpretation of an Archean volcaniclastic sequence, Noranda Region, Quebec, Canada. Canad. J. earth Sci. **15**, 427-436.

Tjia, H.D., 1973a. Palu-Koro fault zone, Sulawesi. Berita Direktorat Geologi 5, p. 3.

-----, 1973b. Displacement patterns of strike-slip faults in Malaysia-Indonesia-Philippines. Geol. Mijnb. **52**, 21-30.

-----, 1981. Examples of young tectonism in Eastern Indonesia. In : Barber, A.J. & Wiryosujono, S. (eds.) The Geology and Tectonics of Eastern Indonesia. Geol. Res. Dev. Centre, Bandung Spec. Pub. 2, 89-104.

-----, and Zakaria, Th., 1974. Palu-Koro strike-slip fault zone, Central Sulawesi. Sains Malaysiana 3, 67-88.

Tucker, M.E., 1985. Shallow-marine carbonate facies and facies models. In : Brenchley, P.J. & Williams, B.P.J. (eds). Sedimentology : Recent Development and Applied Aspects. Geol. Soc. London., 147-172.

Twenhofel, W.H., 1937. Terminology of the fine-grained mechanical sediments : Exhibit F - report of Committee on Sedimentation 1936-1937. Nat. Amer. Assoc. Petrol. Geol. Geog., 81-104.

Umgrove, J.H.F., 1938. Geological history of the East Indies. Am. Assoc. Petrol. Geol. Bull. 22, 1-70.

Untung, M. & Barlow, B.C., 1981 The gravity field in Eastern Indonesia. In : Barber, A.J. & Wiryosujono, S. (eds.) The Geology and Tectonics of Eastern Indonesia. Geol. Res. Dev. Centre, Bandung Spec. Pub. 2, 53-64.

Uyeda, S., 1981. Subduction zones and back arc basins - A review. Geol. Rundschu. **70**, 552-569.

Vail, P.R., Mitchen, R.M.Jr and Thompson, S., 1977a. Seismic stratigraphy and global changes of sea level, part 3 : relative changes of sea level from coastal onlap. In : Seismic stratigraphy - application to hydrocarbon exploration. AAPG. Memoir **26**, 63-81.

-----,-----,-----, 1977b. Seismic stratigraphy and global changes of sea level, part 4 : Global cycles of relative changes of sea level. In : Seismic stratigraphy - application to hydrocarbon explorasion. AAPG. Memoir 26, 83-97.

----- and Todd, R.G., 1981. Northern North Sea Jurassic unconformities, chronostratigraphy and sea-level changes from seismic stratigraphy. In: Illing, L.V. & Hobson, G.D. (eds.) Petroleum geology of the continental shelf of northwest Europe, London, Institute of Petroleum, 216-235.

-----, Hardenbol, J. & Todd, R.G., 1984. Jurassic unconformities, chronostratigraphy and sea-level changes from seismic stratigraphy and biostratigraphy. In: Interregional Unconformities and Hydrocarbon Accumulation. AAPG Memoir, 36, 129-144.

Van Andell, Tj. H., 1975. Mesozoic/Cenozoic calcite compensation depth and the global distribution of calcareous sediments : Earth Planet. Sci. Lett. **26**, 187-194.

Van Houten, 1974. Northern Alpine molasse and similar Cenozoic sequences of southern Europe : In : Dott, R.H.Jr and Shaver, R.H. (eds.) Modern and Ancient Geosaynclinal Sedimentation. SEPM. Spec. Pub. 19, 260-273.

Veevers, J.J., 1976. Western margin of Australia : a Mesozoic analog of the East African rift system. Geology **4**, 713-717.

-----, 1977a. Rifted arc basins and post-breakup rim basins on passive continental margins. Tectonophysics 41, T1-T5.

-----, 1977b. Models of the evolution of the eastern Indian Ocean. In : Heirtzler, J.R., Bolli, H.M., Davies, T.A., Saunders, J.B. & Sclater, J.G. Indoan Ocean Geology and Biostratigraphy. Amer. Geophys. Mem., p. 151-163.

----- & Cotterill, D., 1978. Western margin of Australia : evolution of a rifted arc system. Geol. Soc. Am. Bull. **89**, 337-355.

Verbeck, R.O.M., 1908. Molukken - Verslag, Geologische verkenningstochten in het oostelijk gedeelte van den Nederlandsche Oost-Indischen Archipelago. Mijn. Nederl. Ind. jaarb. 31e, 106-114.

Visser,W.A. & Hermes, J.J., 1962. Geological results of the exploration for oil in Netherlands New Guinea. Verh. Konink. Nederlands. Geol. Mijnb. Genoot. Verh. Geol. Ser. **20**, 265 pp.

Von Kutassy, 1934. In : Bemmelen, R.W. Van (1949) The Geology of Indonesia **Vol.1A**, p. 397.

Walker, R.G., 1975a. Conglomerate : sedimentary structure and facies models. In : Harms, J.C., Southard, J.B., Spearing, D.R. & Walker, R.G. Depositional Environments as Interpreted from Primary Sedimentary Structures and Stratification Sequence. SEPM. Short Course Notes,Tulsa No.2.

-----, 1975b. Generalised facies models for resedimented conglomerates of turbidite association : Geol. Soc. Am. Bull. **86**, 737-748.

-----, 1977. Deposition of Upper Mesozoic resedimented conglomerates and associated turbidites in southwestern Oregon. Geol. Soc. Am. Bull. **88**, 273-285.

-----, 1978. Deep-water sandstone facies and ancient submarine fans : Models for exploration for stratigraphic traps. AAPG. Bull. **62**, 932-966.

-----, 1979. Facies Models.Geol. Assoc. Canad. Geosci. Canad. Ser.**1**, 211 pp.

-----, and Mutti, E., 1973. Turbidite Facies and Facies associations. In : Middleton, G.V. & Bouma, A.H. (eds.) Turbidites and Deep-water Sedimentation. SEPM. Short Course Notes, Anaheim, 119-158.

Wanner, J., 1910. Blitrage zur geologie des ostarmes der ensel Celebe. Unpub. Dies. Jahrb.

Warris, B.J., 1973. Plate tectonics and the evolution of the Timor Sea, northwest Australia. Aust. Petrol. Expl. Assoc. J. **13**, 13-18.

Weiner, R.J., 1978. Deltaic and shallow marine sandstones : sedimentation, tectonics and petroleum occurrences. AAPG. Contin. Educ. Course Note Ser. **2**.

Weissel, J.K., 1980. Evidence for Eocene oceanic crust in the Celebes basin. In : Hayes, D.E. (ed.) The Tectonic and Geologic Evolution of Southeast Asia Seas and Islands. AGU. Geophys. Monogr. **23**, 37-48.

Welland, M.J.P. & Mitchell, A.H.G., 1977. Emplacement of the Oman Ophiolite : a mechanism related to subduction and collision. Bull. Geol. Soc. Am. **88**, 1081-1088.

Wentworth, C.K., 1922. A scale of grade and class terms for clastic sediments. J. Geol. **30**, 377-392.

Westermann, G.E.G., 1973. The Late Triassic Bivalve Monotis. In : Hallam, A. (ed.) Atlas of Palaeobiogeography. New York : Elsevier, p. 251-258.

Williams, H., 1971. Mafic-ultramafic complexes in western

SOME SEDIMENTOLOGIC ASPECTS OF MESOZOIC ROCKS IN EASTERN SULAWESI, INDONESIA. **A3**

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ABSTRACT

Geologic mapping in 1979-1981 of the Malili, Bungku and Poso quadrangles at a scale of 1:250.000 indicates the occurrence of two distinct Mesozoic sedimentary successions in eastern Sulawesi. The first succession is composed of clastic limestone and shale (Tinala Formation) and quartzose arenite and dark-grey shale with coal lenses (Nanaka Formation). These two formations are of continental-shelf deposits, and are together more than 3000 m thick. The second succession is composed of alternations of calcilutite and chert with radiolaria (Tetambahu, Lamusa and Matano Formations), which are all deep-sea origin. The differing lithologies of these two succession reflect their deposition in two separate palaeobasins. Their present juxtaposition occurred after Middle Miocene time, due to underthrusting of the westward-moving Banggai-Sula minicontinent beneath rocks of what is herein called the Eastern Sulawesi Ophiolite Belt.

Oil seeps occur along faults that separate the Tinala and Nanaka Formations from rocks of the Eastern Sulawesi Ophiolite Belt. Lithologies of these two formations suggests that they provide both the source and reservoir of this oil. The emplacement of the ophiolite which now ove lies the Mesozoic continentally-derived sediments and overlying Tertiary rocks are also briefly discussed.

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* Geological Research and Development Centre (GRDC)

M E G A C Y C L I C   T U R B I D I T E S *

by

T.O. Simandjuntak**

Two methods of semiquantitative analysis of turbidites were applied to the Upper Devonian and Lower Carboniferous sedimentary succession in the Cobbadah Distric, New South Wales, Australia.

1. The ABC (Proximality) index (P or P_1) of Walker (1967) based on the proportion of beds in the group beginning with divisions A, B, and C (or T_a, T_b and T_c of Bouma respectively).
2. Ratio of sand to shale of Lovell (1970) based on the proportion of individual beds of sand to those of shale.

Data collected from more than a thousand beds from five sections representing three rock units of different age (Upper Devonian Noumea Beds and Lowana Formation, and Lower Carboniferous Tiabundie beds) show that as P_1 (ABC index) decreases, that is, as depositional environments become more distal, bedthickness decreases, erosion becomes less common, amalgamated sand beds dereases, mud flakes (mud chips) become less common, and the abundance of the parallel and cross-lamination.

Similar trend is also clearly shown by the ratio of sand to shale. The ratio of sand to shale gradually decreases towards distal deposition environment.

Closer field examination, particularly of bedthickness and size-grade as well as other internal sedimentary features have led to conclusions that megacyclic turbidites are well developed in the Upper Devonian Noumea Beds and Lowana Formation.

The *normal megacyclic turbidites* consists of three groups :

1. A basal Proximal Exotic group (P_1 = undetermined, sand/shale ratio > 10 : 1).
2. A middle Classical Proximal Turbidite group (P_1 = 50 to 100%, sand/shale ratio : = 5 : 1 to 10 : 1).
3. An Upper Classical Distal Turbidite group (P_1 = less than 50% and sand/shale ratio = less than 5 : 1 and usually 1 : 1 or less).

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** Geological Survey of Indonesia.

The upward passage between one group and the next is usually gradual with a progressive decrease in both bedthickness and coarseness.

The megacycle is termed *abnormal* if one or two groups are missing. Base on the variation of the occurrence of these groups, the abnormal megacyclic turbidites can therefore be subdivided into base-cut-out, truncated and base-cut-out and truncated megacycles.

The origin of megacyclic turbidites apparently due to two main geological factors : tectonic and sedimentation mechanism. A relative high volcanic activity which produces a complete range of size of rocks (from ash to boulder) couples with capability of the *sediment gravity flows* (debris flows, grain flows, fluidised sediment flows and turbidity current) as transportation agent would produce a normal megacyclic turbidites. A relative low volcanic activity would produce an abnormal megacycle. Sediment gravity flow is believed to have been initiated or even accellerated by a relatively high activity of earthquake pre, syn - and immediately after volcanism (or during tectonism).

On the basis of palaeocurrent analysis, palaeobasin analysis, palaeogeographic reconstruction and provenance and directional of pre-venance these turbidites and their associated coarse clastic rocks are interpreted to have been deposited in *a submarine fan* environment within the trench which was fed by multiple *submarine canyon* from the west.

The occurrence of both large and small scales erosional features, as well as thinning upwards sequence of the Upper Devonian Noumea Beds and Lowana Formation indicate that they have been deposited in a depositional *channeled suprafan* (or *mid-fan*)

PLATE 7

- A Para conglomerate (half bottom) grades to pebbly lithic arenite (top) in H Formation.
- B Pebbly arenite (Ta) grades to coarse grained parallel laminated calc arenite very thinly developed on top), in Halang Formation. Note the presence of erosional surface on both top and bottom of the bed.
- C Deformed convolute lamination in Tc layer of Turbidite in Halang Formation. Arrow shows palaeo current direction.
- D Alternating coarse clastic and silty pelitic rocks in Halang Formation. Note development of parallel sided beds of the thinly bedded silty pelitic rocks on bottom, and slightly grading pebbly lithic arenite on top.
- E Development of truncated and base–cut–out sequence forming by parallel laminated coarse arenite (Tb) on bottom and current ripple laminated arenite (Te) on top in Halang Formation.
- F Development of truncated and base–cut–out sequence (Tb–e) in calc arenite of Halang Formation.
- G Similar to E and F in Halang Formation.

Bull Geol. Res. and Dev. Centre no. 2 1979 .

SEDIMENT GRAVITY FLOW DEPOSITS IN PANGANDARAN–CILACAP REGION, SOUTH–WEST JAVA AND THEIR BEARING ON THE TECTONIC DEVELOPMENT OF SOUTHWESTERN INDONESIA

T.O. Simandjuntak

Originally presented to the 7th Annual Meeting of Indon. Assoc. Geol April 1979.

ABSTRACT

The internal and external sedimentary features observed within the volcanogenic sediments of the Lower Miocene Jampang Formation and the alternating silty pelitic and coarse clastic rocks of the Upper Miocene Halang Formation indicate that these successions were deposited in relatively deep environments by mass–movement of sediment gravity flows.

Parallel and current ripple laminations of the silty pelitic beds, the normally grading and the good stratification of the coarser clastic rocks are the most significant internal sedimentary features of the successions. Turbidite features were frequently observed, although they are commonly of the incomplete type of the Bouma sequences. Parallel sided beds of the relatively thinly bedded silty pelitic rocks are observed. Erosional surface and channeling are the most common external features found within the successions.

Lenses of basaltic and andesitic lava (interfingering ?) found within the Jampang Formation as well as the chert fragments occuring within rudites are believed to have been derived from the socalled Jampang–Menoreh non–volcanic outer arc. This arc is also believed to have been formed during and subsequently after the collision of the north–north–easterly moving Indian Plate and the Asian continent in the Tertiary.

INTRODUCTION

This paper deals with the sedimentary succession in the Pangandaran–Cilacap region (Fig.1). The sediments are deposits of sediment gravity flows of Middleton and Hampton (1973). An attempt is also made to reconstruct the paleogeographic history of the region in relation to tectonic evolution of the southwestern part of the Indonesian Archipelago.

The author is concerned with two main geological problems in connection with the deposition and tectonic history of the region:

(1) It is apparent that there is a structural and tectonic discontinuity between the southeastern part of West Java and southwestern part of Central Java. Surface expression of the discontinuity is now marked by the Citanduy river, which is also known as the provincial boundary of the two blocks.

(2) There are two distinctive sedimentary successions which are classified as deposits of sediment gravity flows, that is the Jampang Beds (lower Miocene or older), and the Halang Beds (Upper Miocene). The sedimentary succession deposited between the two formations may be classified as a traction current deposit (e.g. the Pemali Formation and Penanjung Formation of Middle Miocene). These two types of sediments imply that two types or conditions of depositional and tectonic developments had intermittently occurred within the early Neogene. In Early Miocene or Oligo-Miocene the region was unstable, earthquake and volcanism were very active, and had (re-) generated the sediment gravity

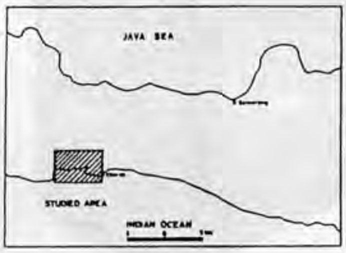


Fig. 1 Index map

flows for transporting and depositing the sediments of the Jampang Beds. In the Middle Miocene as the tectonism and vulcanism temporarily ceased the region was relatively stable and the sea was quiet and traction current would act as a deposition agent for the sediments of the Pemali Formation and Penanjung Formations. In the Upper Miocene, tectonism and volcanism were active again, the basins were unstable, and subsequently followed by the activity of sediment gravity flows for transporting and depositing the sediments of the Halang Beds. The area studied lies in the Southern Mountains of Java which is a physiographic province defined by van Bemmelen (1949) to be a mountainous range underlain by Tertiary volcanics and younger Neogene carbonates. The Jampang Formation discussed herein is a sequence dominated by volcanogenic sediments, distributed along the southernmost, mountainous part of west Java. In other parts the formation is intruded by igneous rocks of granodioritic to dioritic composition. Elsewhere in Central and East Java similar rock units assume other names (van Bemmelen, 1949; Marks, 1957; Bothe, 1929; Sartono, 1960), but they have been collectively known under the (presently obsolete) name "Old Andesite Formation" A number of valleys or depressions which are interpreted as fault zones separate the Southern Mountains into segments. The Citanduy valley is one of these fault zones, which this author assumes as having a considerable tectonic significance. Northeast of the Citanduy Fault Zone another rock unit known as the Halang Formation will be discussed because of its having internal and external properties similar to that of the Jampang Formation.
This paper is the result of observations made during the geological mapping in October – December 1978 which was a part of the 1 : 100.000 scale systematic geological mapping on Jawa. All members of the field team contributed a great deal of knowledge to the author . The geologic map was being compiled as this paper was presented, and will hopefully be published soon. Colleagues who had mapped the adjacent quadrangles have shed much insight to stratigraphic and tectonic knowledge, while senior colleagues have advised to and/or discussed with the author on the regional geology and tectonics.

STRATIGRAPHY

The stratigraphy of the region can be divided into the following units (from base to top) (Fig. 2 and 3).

Jampang Formation	– Oligo–Miocene
Nusakambangan tuff	– Middle Miocene
Pemali Formation	– Middle Miocene
Penanjung beds	– Middle Miocene
Kalipucang limestone	– Middle Miocene
Permisan beds	– Middle Miocene
Halang Formation	– Late Miocene
Kumbang Formation	– Lower Pliocene
Tapak Formation	– Late Pliocene
Iron sands	– Pleistocene–Holocene
Alluvial	– Quaternary

The oldest rock in the region is the Jampang Formation of Late Oligocene to Early Middle

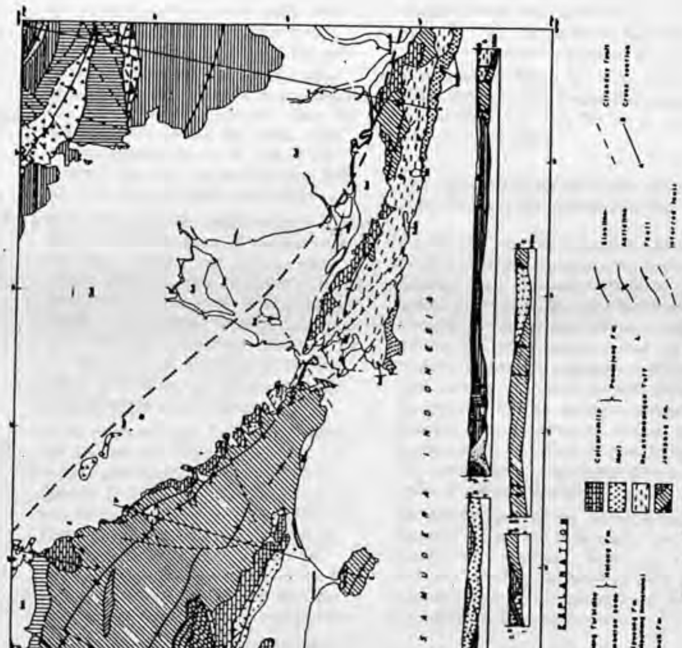


Fig. 2 Geologic map of Pangandaran Region

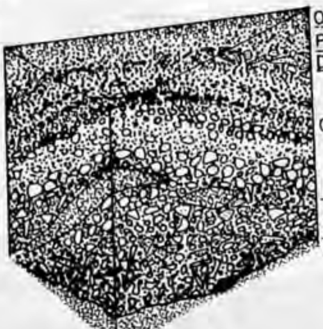


Fig. 4 Block diagram showing a coarse clastic rocks succession. LB=lava basaltic to andesitic, B1=breccia with fragments of basalt (spilitic), andesite, chert, limestone, and mudstone), OC=organised conglomerate, DI=diamictite, PC=paraconglomerate, OC=ortho-conglomerate.

The succession is dominated by volcanogenic sediments such as breccia, tuffs and lava of andesitic and basaltic composition. Subsidiary to these, there are also found rudite, silt, tuff mudstones, and rarely marl intercalations. Basal succession of this unit is not clearly observed in this area. Bemmelen (1949) and Marks (1961) have pointed out that "the lower Jampang Series consist of tuffaceous globigerina marls, tuff sandstones and beds of tuff–breccia of andesitic composition with intercalations of thick limestone lenses containing larger foraminifera indicate a Lower Miocene age".
In the lower part the succession is formed by monolith–breccia and lava of andesite and basalt composition. The breccia has apparently been derived from the lavas. Both lava and breccia often show amygdaloidal features, which are commonly filled by calcite, chalcedony, or zeolite. The latter indicates that the rocks have been subjected to some degree of burial metamorphism. Some lavas could be classified as spilitic rocks; and pillow structure has been observed in a weathered exposure.

On top of this, there is a significant lensoidal breccia which contains limestone and cherty argillite subsidiary to volcanic fragments, some of which show spilitic appearance. On closer field examination the breccia can be classified as disorganised conglomerate with components ranging in size from pebble to nearly 1 metre, subrounded to angular in shape, and randomly distributed throughout the bedding. The base and upper contacts show erosional features, and therefore the rock has been deposited in submarine channeling. Breccia and lavas are believed to have been derived from the socalled Jampang–Menoreh High, which was developed during and subsequently after Tertiary collision. This will be discussed in more detail in the Tectonic and Depositional Development in the following section.

In the upper part the succession consists of alternating coarse and fine–grained volcanogenic clastic rocks with marly tuff intercalations containing foraminifera of Lower Miocene age. The rocks often show sedimentary features of both internal and external structures. The most common internal sedimentary structure includes graded bedding, laminae 1 within the coarser clastic rocks, and laminae 2, current ripple lamination, microcrossbedding, convolute lamination and slump within pelitic rocks. Diffuse lamination is rarely observed in arenites and pebbly arenites. Current foresets have also been observed within some arenite and pebbly arenite bedding. External sedimentary structures such as erosional surfaces and channeling are frequently observed, particularly on the base of arenites, pebbly arenites, diamictite, para–and ortho–conglomerates. These structures have been observed in all scales, from mesoscopic to megascopic size (less than 1 metre up to several kilometres long) such as that in Nusakambangan and S. Cipakuhaji (Plate 1 E, Plate 2 B and F). Load casting has also been frequently observed on the base of the coarser clastic rocks (Plate 2 D). Flutecasts are rarely observed (Plate 6 E), possibly due to the feldspathic (or volcanogenic) nature of most of the coarse clastic rocks. Amalgamated beds have been observed, where two or more layers of coarse clastic rocks (arenites, pebbly arenites) were jointed (associated) together without any silty pelitic layer(s) between them. Parallel sided beds are the most significant external sedimentary features have been observed within the relatively thinly bedded silty pelitic rocks and fine arenites (plate 2 A, Plate 4 D). Pullapart structure and sand injection have also been observed within these rocks.

Data collected on the palaeocurrent indicators show that the lower part of the succession were largely derived from south or southwesterly direction, while the upper part were mostly derived from north or northwesterly direction and subsidiary materials have come from south and southeasterly direction. This will match later on discussion on the direction of provenance, and interpretation of palaeogeography of the region.

In the Nusakambangan Tuffs similar sedimentary features were frequently observed, particularly grading (Plate 1 F), stratification (laminae 1) (Plate 1 A, D, F, Plate 4 D) current foreset, convolute lamination (Plate 1 C, Plate 4 A), parellel lamination (laminae 2) and current ripple lamination (Plate 4 B, E). These structures and nature of ash fall tuffs of most succession indicate that they have been deposited in marine environment.

The most common turbidity features observed within Jampang Formation and Nusakambangan Tuffs are grading (Ta) in arenites, pebbly arenites and conglomerates, parallel lamination (Tb–laminae 1)of arenites and sometimes even in pebbly arenites, convolute lamination or current ripple lamination and micro cross bedding in fine–to medium–grained arenites or siltstones, and parallel lamination (laminae 2) of silty pelitic rocks (Plate 1 F) Pelitic (Te) is rarely observed. These structures are usually developed in an incomplete sequence of Bouma (1962), as truncated (Plate 2 B, F, Plate 3 G) or base–cut or truncated and base–cutout sequence (Plat 2 D). The complete sequences appear to have not been developed in the succession On closer field examination it can be roughly estimated that the ratio of coarse clastic rocks to silty pelitic rocks is of about 75 : 25. This indicates that succession were deposited in proximal depositior environment of Walker and Mutti (1973) a Haner (1971). More detailed discussion an interpretation dealt with this can be read later on in this paper.

THE PEMALI FORMATION AND PENANJUNG FORMATION (Middle Miocene)

The Pemali Formation exposed in the regi to the east of the Citanduy valley is dominated by marly mudstones with thinly bedded limestone lenses, while the Penanju Formation in the Pangandaran area west the Citanduy valley consists of marl and bioclastic limestone (calcarenite) in about equal amounts. The rocks are well bedded but internally they are lacking in sediment structures. Most beddings are massive, although laminae may occur in some beddin The coarser clastic rocks usually show a randomly oriented fragments.

The data indicate that the succession were deposited in open and quiet marine enviro ment by traction current. Sedimentary structures as well as the lithology make th succession quite distinct and different from the underlying Jampang Formation.

THE HALANG FORMATION (Late Miocen to Early Pliocene)

As has been mentioned the Halang Formati which is much thicker in the region to the east of the Citanduy valley than in the we consists of conglomerates, diamictite, pebbl arenites, and calcarenite, marls, calcareous silty pelitic rocks and mudstone.

Sedimentary structures developed in this un is about similar to those found in the Jampa Formation, although in the Halang Formation turbidity features are more significant and wel

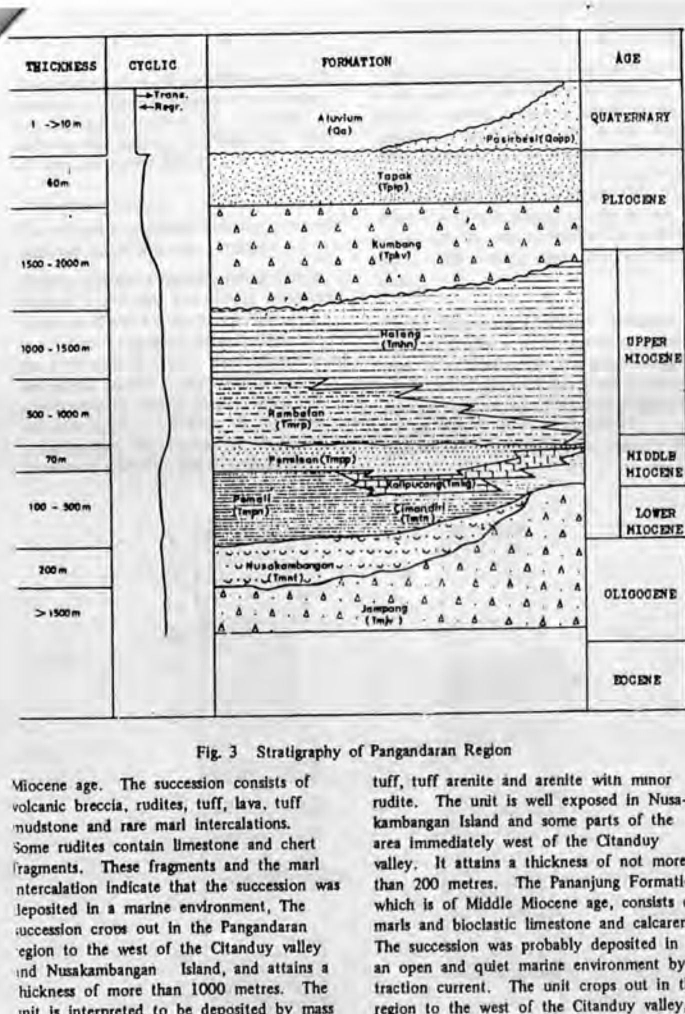


Fig. 3 Stratigraphy of Pangandaran Region

Miocene age. The succession consists of volcanic breccia, rudites, tuff, lava, tuff mudstone and rare marl intercalations. Some rudites contain limestone and chert fragments. These fragments and the marl intercalation indicate that the succession was deposited in a marine environment, The succession crops out in the Pangandaran region to the west of the Citanduy valley and Nusakambangan Island, and attains a thickness of more than 1000 metres. The unit is interpreted to be deposited by mass movement of sediment gravity flows. The overlying Nusakambangan Tuffs of Middle Miocene age, consists dominantly of ashfall tuff, tuff arenite and arenite with minor rudite. The unit is well exposed in Nusakambangan Island and some parts of the area immediately west of the Citanduy valley. It attains a thickness of not more than 200 metres. The Pananjung Formation which is of Middle Miocene age, consists of marls and bioclastic limestone and calcarenite The succession was probably deposited in an open and quiet marine environment by traction current. The unit crops out in the region to the west of the Citanduy valley, and ranges in thickness from 100 to 500 metres. To the east of the Citanduy valley the equivalent unit is dominated by marly mudstone with thinly bedded calcarenite intercalation (Pemali Formation) with a thickness of 100 metres.

The Kalipucang Limestone consists of reef and bedded limestone and varies ranges in thickness 25 to 100 metres. The reefs apparently started to develop on top of the Jampang Formation and continued to build up in Middle Miocene in or along the eastern and of the rising "Jampang High". Reefs are not found in the region east of the Citanduy valley.

In Nusakambangan, the Permisan Formation which is dominated by arenite and rudite overlies the Nusakambangan Tuffs and attains a thickness up to 100 metres. The age of this unit is Middle Miocene.

The Pemali Formation and the Kalipucang Limestone are overlain by the Halang Formation of Late Miocene to Early Pliocene age. The lower member (Tmrp) consists of rudite, arenite, marly mudstones, while the upper member (Tmhn) is dominated by marls, calcareous mudstone with intercalated calcarenite and rudite. This sequence was probably deposited by sediment gravity flows in a relatively deep–environment, and attains a thickness varying between 500 to 1500 metres. This unit is much thicker in the region east of the Citanduy valley.

A thick volcanogenic sediment sequence overlying unconformably the Halang Formation is named Kumbang Formation of Pliocene age. which consists of breccia, lava, and arenite with marl intercalation and intrusions, and reaches a maximum thickness of 1000 metres. The succession was apparently deposited in a marine environment by sediment gravity flows.

The youngest Tertiary sedimentary succession is the Tapak Formation of Late Pliocene age, which consists of arenite with marl intercalations; it attains a thickness of not more than 60 metres, and was deposited in a littoral to neritic environment.

'Iron sand' is a Quaternary beach deposit which contains magnetite up to 45% and limestone up to 25%. It is locally found along the beach in the vicinity of Cilacap and at some places of Nusakambangan beach.

Alluvial deposits extensively cover the Citanduy valley and other low and flatlying areas and consist of muddy silt, arenite and gravel derived from older rocks.

FIELD DATA RECORDS

On the basis of field observations, sedimentary features observed in both the Jampang and Halang Formations can be divided into two types :

(i) Internal sedimentary structures, that is, all structures shown by the arrangement of fragments or particles forming the body (layer, bedding) of the rocks. The structures are genetically formed syndepositionally with the rocks. This type includes normal or reverse grading, laminae 1 of Lombard, 1963 (stratification), thin parallel lamination (laminae 2 of Lombard, 1963), current ripple lamination, dish structure of Walker (1976), micro cross bedding, current foreset, diffuse lamination, mud flakes, arrangement of fossils (Fig. 4).

(ii) external sedimentary structures, which includes features of rocks as a whole body (layer, bedding) of rocks. This includes, features of the base and top (upper) contact of layer or bedding, such as erosional surfaces, channeling, load casting, flute casting, rippsole marking, amalgamated beds, parallel sided beds, etc. The external features can be formed syn–or post–depositionally.

THE JAMPANG FORMATION

As described previously the Jampang Formation crop out in Pangandaran region and Nusakambangan, and is not found in the region to the east of the Citanduy valley. This is apparently controlled by tectonic and basinal development of the region, particularly during deposition of the sediments. This will be separately discussed later on in this paper.

developed than in the Jampang Formation. The presence of alternating coarse clastic rocks and silty pelitic rocks in this succession indicates that they are turbidites (one type of sediment gravity flows).

Sedimentary Features

The internal sedimentary structure commonly observed within this unit, includes :

Grading, usually of normal grading (Ta) in arenites, calcarenites, and even in orthoconglomerates (Plate 5 B, D, F and Plate A, B, C, D). In the presence of mud flakes in the lower part of the Ta, the fragments which are usually parallel to and dipping upstream a indicating by current foreset within the associate layers or beddings. In organised conglomerates, the components which usually consists of volcanics, and of sedimentary origin (limestone, arenites, laminated mudstone, marls) in the base of the beds are graded to pebbly arenite or arenite at the middle and upper part of the layers. Sometimes the lower part is formed by diamictite and then grading upwards to paraconglomerates or pebbly arenite to arenite (Fig. 5). Reverse grading if present usually in beds where only Ta with or without Tb is developed. This is rarely observed in this succession.

Parallel lamination. Laminae I (Lombard, 1963, – parallel lamination with space more than 2 cm between one and other laminae) usually develops as Tb composed by arenites, calcarenites, or even pebbly (calc) arenites (Plate 7 B, C, D, F). Sometimes lamination is disrupted by the presence of

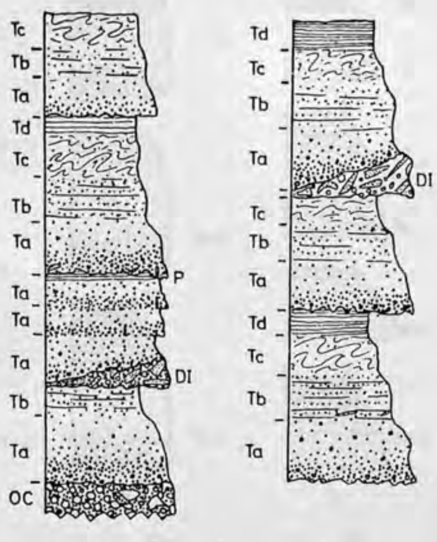


Fig. 5 Development of turbidites in Jampang Beds (A) and in Halang Formation (B) They are mostly of incomplete type.

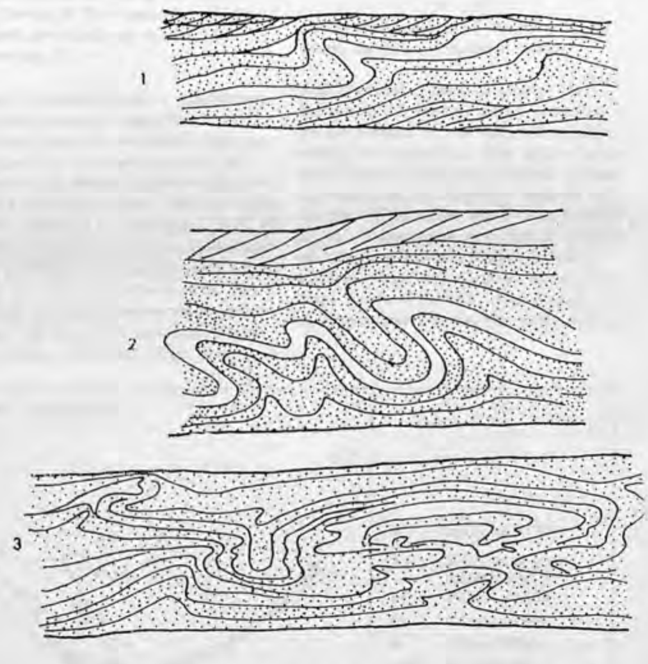


Fig. 6 Tracing of current ripple lamination in T_c layers of turbidites in the study area, showing gradation from weakly distorted current ripple laminae (1) to highly convoluted laminae (3).

Fig. 7 Complete sequence $T_1 = T_{a-e}$

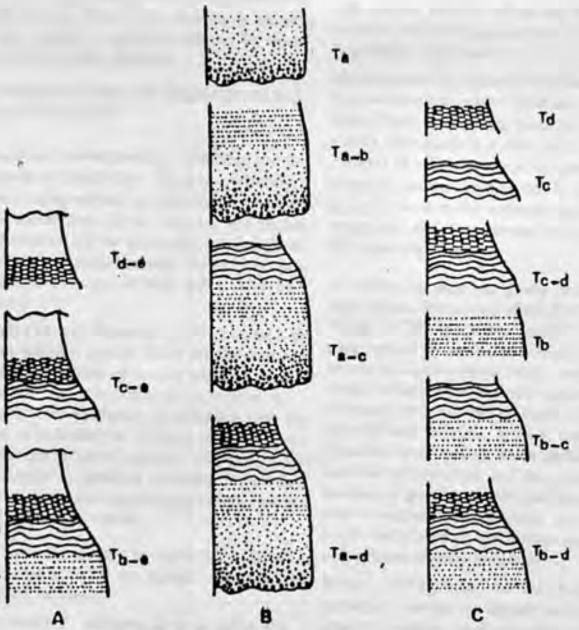


Fig. 8 Configuration of incomplete sequence, base cut-out units (A), truncated units (B), and truncated base cut-out units (C).

mud flakes in the lower part of the Tb. The parallel lamination is marked by a change in amount of mud present within the arenite or pebbly arenite. Individual laminae ranges in thickness from more than 2 cm to 10 cm. Tb is sometimes graded, but in general it is poorly graded or even massive, and mainly occurs in association with graded interval (Ta). Contact of Tb is generally gradational with both the underlying intervals (Ta) and the overlying Tc.

Laminae 2 (Lombard, op.cit – parallel lamination with thickness ranging from few mm to 20 mm), is usually developed in silty pelitic rocks (Td). This lamination can be observed on the basis of differences in color, and has a gradational contact with the underlying Tc. (Plate 6 D, and Plate 7 D, C, F). *Convolute lamination.* Convolute lamination usually develops in Tc intetval, which ranges in grain size from silt to fine arenite. This lamination sometimes contains micro-cross-bedding. On closer field examination some laminae show a gradational shape from undeformed current ripple lamination to deformed current ripple lamination or to convolute lamination (Plate 6 D, F, Plate 7F, G and Fig. 6). In turbidite deposits, convolute lamination can be used to determine palaeocurrent, by measuring direction of the overturned apices of convolute lamination.

The external sedimentary structures found within the Halang Formation include :

Bed form, and bedding continuity. The features of the beds are largely controlled by the presence of erosional surfaces, channeling, load casting and other surface feature: In the Halang Formation, erosional surface and channeling are frequently observed on the base of the coarse clastic rocks (Plate 7B, D, F). Channeling is commonly of mesoscopic size, ranges from a few cm to tens of cm depth, from tens of cm to several meters long. Erosional surfaces often develop on the upper surface of the silty pelitic rocks.

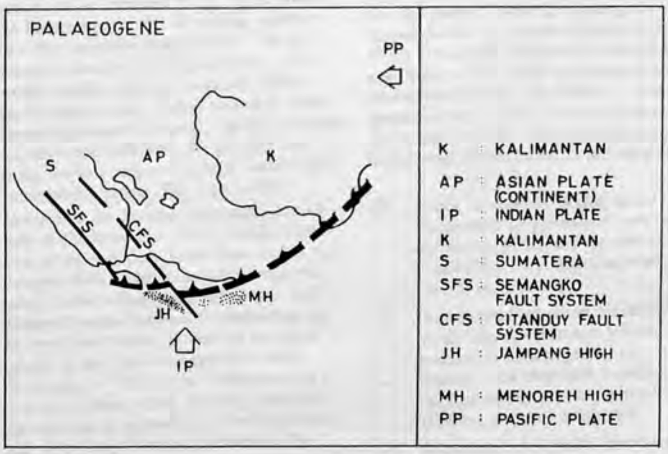


Fig. 9 Development of the Citanduy Fault System

Load casting is also found particularly on the base of arenites underlain by pelitic rocks. It ranges from a few mm up to tens of cm in size (Plate 7 E).

Flute casting is frequently observed on the base of the arenite beds, especially that of calcarenites. It varies in length between 2 cm to 25 cm and indicate a paleocurrent direction of a similar trend shown by current foresets.

Parallel sided beds is the most common external sedimentary structure observed within the relatively thinly bedded silty pelite and fine arenite (Plate 7 C). Bedding is generally quite regular in thickness and can be traced for considerable distances.

INTERPRETATION OF SEDIMENTOLOGICAL PROCESSES

Based on the sedimentary structures and lithological distribution described previously, some interpretation of sedimentary processes and mechanisms of the Jampang and Halang Fôrmations will be attempted. Relations of sedimentary processes with depositional environment and type of basin will also be discussed.

Middleton and Hampton (1973) proposed the term *sediment gravity flows* for the transportation mechanism of arenite and coarser sediments into deep water via submarine canyons. The mechanism is a general term for flow of sediment or sediment–fluid mixtures under the action of gravity. The gravity flow mechanism is classified according to the dominant sediment–support mechanism involved and defined as follows :

(i) *Turbidity current* in which the sediment is supported by the upward component of fluid turbulence.

(ii) *Fluidised sediment flows* in which the sediment is supported by the upward flow of fluid escaping between the grains as the grains are settled out by gravity.

(iii) Grain flows in which the sediment is supported by direct grain interactions (collisions or close approaches).

(iv) *Debris flows* in which the sediment is supported by a "matrix" that is, by a mixture of interstitial fluid and fine sediments, which has a finite yield strength.

Sedimentary gravity flow is distinguished from gravity sliding or slumping on the basis of the degree of internal deformation, which is intensive in flows, slight in sliding and intermediate in slump (Verness, 1958, and Dott, 1963). In slides, large block or slab of material move on a few relatively well defined slippage planes, but in slumps the material may break up into many blocks and is generally deformed.

Middleton and Hampton (op.cit) also explained that sedimentary gravity flow may be distinguised from fluid gravity flow; in a fluid gravity flow (such as a river or an occean current) the fluid is moved by gravity and drove the sediment along, but in sediment gravity flow it is the sediment that is moved by gravity, and the sediment motion moves the interstitial fluid.

In relation to sediment gravity flow mechanism above, Walker and Mutti (1973) and Walker (1976) and Hampton (1972) recognised several facies in the turbidites and the associated coarse–clastic rocks, based on several criteria, including size–grade, bed thickness, ratio of sand to shale, bedding regularity, sole markings, internal sedimentary structures and textures which include conglomerate pebble fabric and the presence or absence of grading, massive bedding of sand with or without dish structure, and variations in the occurrence of the Bouma sequence and palaeoecological indicators.

Walker (1976) pointed out that because a turbidite is simply the deposit of a turbidity current, turbidites would therefore be found in any environment where turbidity current operate. The environments included lakes and reservoirs, delta fronts, continental shelves, and importantly the deeper oceanic basins, (continental rises and abyssal plains). Walker (op.cit) then pointed out that the classical turbidites can be characterised by three main features: firstly, the beds tend to be laterally extensive (hundreds of meters); secondly, they tend to be parallel sided and vary little in thickness laterally (hundreds of meters); and thirdly, it is reasonable to use Bouma model for their description and interpretation. In addition to that, Walker (op.cit) has suggested four facies coarse clastic rocks associated with turbidites, they are (a) massive sandstone, (b) pebbly sandstone, (c) clast supported conglomerates and (d) chaotic matrix–supported sandstone and conglomerates. (Figs. 7, 8).

All these parameters will be discussed in particularly relation with the occurrence of them in Jampang and Halang Formation in the study area.

Massive arenite facies which is comprised of thick arenite beds in which grading is normally poorly developed. A typical sequence of beds would be measured as T_a, T_a, T_a using the Bouma model, with or without structure, very thin mudstone intercalations. The sequence is therefore susually truncated. The massive arenite units are commonly not as parallel sided as the classical turbidites, while channeling is more common, and one flow may cut down and weld onto the previous one giving rise to a series of multiple arenite beds (amalgamated beds). This type is often observed in the Jampang Formation, particularly in the lower part, but in Halang Formation is less commonly found than that in the Jampang Beds. Dish structure is more commonly found within the arenite beds. Bed thickness ranges from ten of centimetres up to over three meters, or even up to ten of meters in the case of amalgamated beds. Mechanism of this arenite is interpreted as a turbidity current which normally maintains its sand load in suspension by fluid turbulence, can pass trough a stage of fluidised flow during the final few seconds or minutes of flow immediately preceding deposition (Fig. 5).

Pebbly Sandstone Facies. This facies usually cannot be described by using the Bouma model. They are often graded, as indicated by a decrease in pebble size towards the top, or an increase in the amount of pebbles towards the base of the beds. Stratifications are rather coarse, crude and horizontal or sometimes even cross–stratification of trough to planar–tabular type are present (Walker, 1976) (Fig. 4).

Pebbly arenite beds are commonly channeled, laterally discontinuous and are usually interbedded with coarse to very coarse arenites or conglomerates or diamictites. Interbedded with mudstones are rarely found.

This type was frequently observed in both the Jampang and Halang Formations and ranges in thickness from a few to tens of centimeters.

Clast supported conglomerate facies. Walker (op. cit) has proposed some generalised "Bouma like" models for sedimented conglomerates, which were based on published data and on his own observations in California, and also the present author has observed in Tamworth Belt in Australia (Simanjuntak, 1977). Descriptive sedimentary features that contribute to the models are preferred clast fabrics, stratification and inverse or normal grading or both. Based on the occurrence and absence of these features, three models for the conglomerate facies can be distinguished

(1) *Inverse to normally graded model* . characterised by presence of both inverse and normal grading and the absence of stratification. A preferred orientation of clasts is often present.

(2) *Graded–stratified model* characterised by the absence of inverse grading and presence of normal grading and stratification. The clasts size is smaller than the first model, and they generally exhibit a preferred orientation.

(3) *Disorganised–bed model* characterised by the absence of normal and inverse grading and stratification. Few of the

nglomerates exhibit a preferred fabric.

In both Jampang and Halang Formations, these three models were often observed. Normally graded beds and absence of stratification is often observed than inverse grading. Sometimes the rocks show a preferred orientation of clasts. Graded–stratified model is frequently observed in both successions, and is classified as organised conglomerate. In the Jampang Formation preferred orientation of clasts is rarely observed, but in the Halang Formation a preferred fabric is more frequently observed. Disorganised–bed model is more commonly found the Jampang Formation than in the Halang Formation. The clasts of the conglomerates are generally randomly oriented, poorly sorted, subangular to rounded in shape, and ranges in size from pebble up to 30 cm.

In general the preferred fabric normally taken the form of clasts with their long axes (a) parallel to the palaeoflow direction and imbricating upstream. This can be explained, experimentally or naturally, as the clasts moved in the bed without rolling. The preferred orientation of clasts is different from that of fluvial deposits, where pebbles and cobbles are rolled on the bed, which usually orients the long axes (a) transverse to the flow direction and the intermediate axis (b) imbri cate upstream.

The preferred clasts fabric with the long axes (a) parellel and imbricated upstream in this conglomerate facies are believed to be deposit of either flows or result of dispersion of clasts in fluid the bed (fluidised flows). Mass movement in which clast are free to move relative to each other do not produce abundant graded bedding, stratification or cross–stratification. Walker (op. cit) suggested that these clasts were supported above the bed in turbulent flow. The supporting mechanism may been partly fluid turbulence and partly clasts collision.

Chaotic–Matrix–supported Pebbly Arenite and Conglomerate Facies.

Walker (op.cit) described two different types of deposits in this facies; the conglomerates and pebbly sandstones that have an abundant muddy matrix, and possibly show basal inverse grading and preferred clast alignment. They are simply deposited by subaqueous debris flows. Because the larger clasts in a debris flow are maintained above the bed by the strength of the debris flow matrix, the deposit commonly large blocks projecting up above the bed, or even resting almost entirely on top the bed. The deposits show no internal evidence of slumping.

The second type of deposit commonly shows evidence of slumping, and represents the mixing of sediment within the depositional basin by post depositional slumping. They can range all the way from very cohesive slumps involving many beds, to very watery slumps generated by deposition of coarse sedimentation top of wet, poorly consolidated clays. The latter process give rise to the pebbly mudstone (Walker, op.cit), diamictite (Flint, and Sanders, 1960). They are deposited by subaqueous debris flows, and Walker (op.cit) has pointed out that inasmuch as subaqueous debris flows and slumps require greater slopes than the classical turbidity currents, the chaotic facies is most abundant at the foot of the slope in the basin. Crowell (1957) suggested that the indurated clasts in pebbly mudstone may have been transported into the depositional area in the first place by vigorous turbidity currents.

These two types were observed in both the Jampang and Halang Formations, and usually in association with the disorganised conglomerates. They often contain subrounded to angular indurated clasts in a homogeneous matrix of muddy sand, or calcareous muddy sand, or calcareous muddy sand. Some of the sedimentary fragments, particularly the larger ones, show evidence of plastic deformation either as fluid (matrix) or as rigid fragments (pebbles).

Turbidite Features

The turbidite deposits in both the Jampang Beds and Halang Formation as described previously, were developed mostly in incomplete sequences. Almost all of the fine characteristic sedimentary features ascribed to turbidite deposition by Bouma (1962) can be seen in the Jampang and Halang Formations. The base of the beds, particularly of the coarser clastic rocks is invariably sharp and there channeling and load cast castings are often found. The arenites commonly grade upwards into fine grained argillaceous mudstone or mudstone at the top. The beds are rarely observed to pinch out, although in some places two or more separate beds can be traced laterally into a single amalgamated unit. In general individual beds, particularly the finer grained succession are laterally persistent over considerable distances.

Sole markings, a common feature of most turbidite sequences are present although are not well preserved in the Halang Formation and some in the Jampang Formation. In many cases low burial metamorphism has effectively welded the underlying mudstone to the base of the overlying arenite beds which would prevent the exposure of the sole of those beds. The features are therefore best seen in a cross section. This includes small scale, undeformed scours and channel fillings. On closer field examination, the occurrence of turbidites is commonly in incomplete sequences, such as truncated sequence forming by T_a, $T_{a\text{-}b}$, $T_{a\text{-}c}$, and $T_{a\text{-}d}$; base–cut–out : $T_{b\text{-}e}$, $T_{c\text{-}e}$, $T_{d\text{-}e}$, T_e; and combination of truncated and base-cut and base-cut-out : T_b; $T_{b\text{-}c}$, T_c, $T_{c\text{-}d}$, T_d. While the complete sequence is rarely observed (Fig. 8).

The occurrence of several $T_{a\text{-}d}$ sequence in, particularly, the Halang Formation makes it possible that a very thin pelitic interval may be found at the top of T_d interval, as Walker (1967) stated that it is often difficult to separate T_e from T_d intervals, especially in weathered exposures. The T_{a-d} sequence observed in the study area have been consequently treated or at least assumed to represent a complete sequence (T_1).

To complete a discussion of turbidites it is necessary to see some significance of the turbidite features occurring in the successi This would include the significance of gra characteristics, bed forms, occurrence of tl mud–flake conglomerate, and pebbly mud stone, amalgamation, occurrence of lamina and cross–lamination, load casting.

Graded bedding.

Kuenen and Migliorini (1950) visualised a turbidity current as a current with a large amount of suspended sediment, of which coarse material is concentrated near the hea of the current and close to the bottom. Presumably, there is lateral and vertical gradi towards finer sized material away from this high–concentration region. At a particula place, material is deposited by parts of the current, and this process results in a vertic grading.

Kuenen and Menard (1952) attributed poo grading near the source in their experimen to rapid deposition. Middleton (1966) sug gested that, in the initial stages, sediment i transferred from the body of the current t the head. Such a current can give rise to poorly grading beds; this is called immatur current, (Walker, 1965).

Middleton (1967) in this experimental wor recognised two types of turbidity currents low–concentration flow and high–concentr tion flows. Within the low–concentration flow, deposition is slow and particle–by–particle, and the whole size distribution shifts to finer size progressively from the t of the beds. This gives rise to normal grad or distribution grading. This type of gradi is characteristic of graded deposited layer–by–layer from relatively dilute experiment turbidity currents (Middleton and Hampton. 1973). In the high–concentration flows there is a sudden deposition of a "quick" b which is sheared extensively. Kuenen (196 has also observed a similar quick bed which he called a "traction carper", and then suggested that the beds formed by high-conce tration flows may be normally graded if sus pended material in the turbidity currents is normally graded. This quick bed is thought to be relatively homogeneous but the materials work their way through to the bottom and "freezing" of such a bed gives rise to coarse tail grading. This grading is probably due to the decay of initial turbulence, in part to decreasing competency in the tail of the turbidity current and part to the concentration coarse material in the head of the head of the current by high-concentration flows. Grain sizes decreasing away from the source have been attributed to a decrease in current velocity and a consequent decrease in the capacity of the current to carry coarse grains. This leads to a progressive loss of coarse materials (Kuenen and Migliorini, 1950, and Middleton, 1967). In all their experiments on turbidy currents, deposition of sediment started just after the formation of the turbidity current. This, could be due to the small size of the apparatus and the high setting velocity of the sediment used (Prakash and Middleton, 1970). In nature, a turbidity current may carry its load of sediment from long distances, possibly in autosuspension, before it starts to deposit its load. The experimental results therefore, may not explain all the features observed in natural beds, particularly in the case of longitudinal changes in grading characteristics.

In the Jampang Beds and Halang Formation most of the arenites and pebbly arenites show good grading. The natural turbidity currents apparently travelled long distances and probably developed a very good longitudinal and vertical grading leading to deposition of beds with good grading in the proximal region.

Bed Forms. The regularity of both bed form and thickness gradually increase towards the distal region. This is due to the decrease of erosion activity towards distal environments. The regularity of bed forms is difficult to quantify and consequently data on them can not be treated statistically. In the field, particularly in the Halang Formation, it is quite clear that the regularity of both bed form and thickness is significantly pronounced within the succession dominated by silty pelitic rocks, or alternating beds of arenites and mudstone with thickness ranging from 3 cm up to less than 25 cm. They also show remarkable persistence in bed form, and individual beds can be traced a considerable distance along strike. However, the regularity of both bed form and thickness observed in the coarse clastic rocks is different, particularly in the Jampang Formation. This could be due to the occurrence of erosional, channeling, load castings on the base of the coarse clastic rocks.

Occurrence of mud flakes conglomerate and pebbly mudstone

In both the Jampang and Halang Formations mud flakes conglomerates are commonly found near the base (although some in the middle) of the arenite beds. Lovell (1970) showed that the mud flakes conglomerates are more commonly found in proximal environment. The mudflakes generally range in size from a few mm up to 20 cm, although some, but rarely, up to 50 cm. They usually occur in bed thickness not less than 10 cm and more frequent in the thicker beds.

Amalgamation Beds.

Amalgamation features have been observed in both the Jampang and Halang Formations. Two or more arenite beds have been welded together by cutting out the original interturbidite material. Kuenen and Menard (1952) suggested that amalgamated beds may simply be the result of an initial lack of mudstone deposition or the erosion and removal of mudstone by the currents depositing the arenites, while Walker (1967) suggested that the increase in amalgamated beds probably related to the increasing erosive ability of high flow regime turbidity currents.

Occurrence of laminae and cross-laminae.

Although laminae I (T_b interval) and cross-laminae (T_c interval) are fundamental units of the Bouma sequence, they do not always occur in turbidites. Sander (1965) pointed out that it is possible that a turbidity current may slow down and rapidly pass trough the range of current velocities suitable for the deformation of laminae and current ripples. The resultant turbidites will consist of T(a–e) sequence with no internal sedimentary structures. This is called a real turbidite, but is rarely observed in the Jampang and Halang Formations.

Load casts

Load casts are sole markings most frequently observed within the Jampang and Halang Formation. The casts vary in length from 2 cm to 20 cm and are mostly oval in shape. They are generally asymetrical in form, probably due to horizontal movements and ratio of the viscosities of the two layers involved. Anketell, et.al (1970) stated that the form of this structure depends on the ratio of viscosities of the two layers; if the viscosities are almost equal the loading will produce symmetrical sinuous deformation, but if the viscosity of the underlying layer is much greater than that of the overlying sand layer, broad, rounded loads of sand will occur with sharp crested flames of mud extending upward between them. The latter is the most frequent form developed within the turbidite succession in the Pangandaran–Cilacap region.

Normally load casts can be formed syndepositionally if material of a sand layer sinks down locally into the underlying layer.

This is due to the occurrence of a reversed density gradient, that is, when the sand bed is denser than mud on which it is deposited and when the viscosity and strength of the sand and mud are sufficiently low to permit deformation of the interface between them. If the structures sink down further into the underlying sand material without additional supplies of sand, the sides of the load cast introvert near the contact with the overlaying sand. Such structures have been produced experimentally by Kuenen (1958). The loads seem to be localised by initial irregularities in the mud surface or in the sand deposited on the mud surfaces. Consequently many flutes may also show loading and load-casted ripples may if a sand ripple sinks down into mu as it is being formed. The materials i pockets in arenite generally are a litt than that found in the overlying ar layer itself.

The load casts are generally develop base of the arenite and pebbly areni conglomerate beds in both Halang F Jampang Beds.

The Condition of the source area a depositional environment

As has been described the developm turbidite successions in both the Ja and Halang Formations respectively rupted by a traction current deposi Middle Miocene age the Pemali For and the Pananjung Formation.

This leads to the conclusion, that formation of these two types of se were largely controlled by two mai ogical factors; 1) tectonics and 2) tional process (agents) and environ

Tectonics

As will be broadly discussed in the section in this paper, the tectonic factor is closely related to source a tion of provenance), initiation of tl and instrumentation of depositional nism, which is also indirectly influc the second factor above.

The melange complexes in Ciletuh (? & Suparka, 1977), Lok Ulo (Suken Peg. Meratus (Bemmelen, 1949), is indication of sediment bearing in it. gene sediments, such as Bayah Beds (Bemmelen, 1949), Cicarucup Beds (Miocene, Bemmelen, 1949), Ciletuh (Eocene, Martojoyo & Suparka, 197 Sukamto, 1975), in West Java; Sou Mts (Eocene, Harloff, 1929–1930), Serayu Mts (Eocene, Tan Sin Hok Bemmelen, 1949), Wungkal Format

Eocene, Bothe, A.ch.D., 1929) in Central Java, and Mahakam Formation (Lower Eocene, Bemmelen, 1949), Kihamhalo Formation (Lower Eocene, Bemmelen, 1949), Karan Formation (Eocene, Bemmelen, 1949 and Marks, 1961) and Taballar Formation (Vlerk, van der, 1931) in Kalimantan are mostly of marine sediments although some are paralic/neritic, and contain materials of mixed continental and oceanic origin. In several they are unconformably underlain by the Mesozoic rocks (basement ?). These rocks are believed to have been deposited within the fore arc basin in the sense of the plate tectonic theory.

The Neogene sediments, such as the Jampang Formation in the study area is dominated by volcanogenic sediments, particularly in the middle to upper part of the succession, while in the lower part it is formed by sediments containing cherty argillite, limestone and spilitic rocks subsidiary to andesitic and basaltic constituents which are often associated with basaltic and andesitic lava. These rocks are thought to have been derived from the socalled Jampang Menoreh nonvolcanic outer (?) are, which is developed as a submarine high during and subsequently after the Tertiary collision between the Indian Ocean plate. The volcanic chain apparently had started to be active in Early Miocene and become the provenence of the volcanogenic sediments in the middle and upper of the Jampang Formation. Lithological and sedimentary differences of these two units have already been described. The Jampang Formation is therefore bimodal in the provenance sense; partly (particularly the lower part) is derived from south–southwesterly direction, while mostly the middle and upper part have been derived from north to northwest or northeasterly direction.

Depositional agents and environments

Since the Jampang–Menoreh High was formed, in Late Paleogene the upper layer or the oceanic crust (e.g. lava and associated ribbon chert and limestone) had been fragmentally faulted while lavas were brecciated to become monolith breccia in association with breccia that contains chert, limestone fragment forming the lower part of the Jampang Formation. These rocks are thought to have been deposited in both the upper slope of a trench and arc trench gap (inter deep) basins. The basins are both deep and open marine. During the late stage of collision earthquake and volcanism were both either spontaneously or intermittently active. Earthquake shocks then generated and initiated the occurence of sediment gravity flows which become agent for transporting and depositing the Jampang Formation. The higher intensity of earthquake and volcanism would produce a higher activity of sediment gravity flows and huge volcanic products and hence would produce a thicker sedimentary succession (e.g. Jampang Formation of more than 1000 m thick).

The occurrence of alternating volcanogenic sediments and the clastic rocks in the middle and upper of the Jampang Beds indicate that volcanism was intermittently active; each activity provided volcanic materials for each volcanogenic unit above.

As the Jampang–Menoreh High rising up with time to become the socalled Jampang–Menoreh nonvolcanic outer arc, the foredeep (trench) separated from the inter deep (arc–trench gap basin). The basin then became shallower with time, due to intensive and extensive sedimentary filling in the Lower Miocene (e.g. Jampang Beds and Nusakambangan Tuffs). The Nusakambangan Tuffs which are dominated by ash–fall tuffs are believed to have been deposited during the late stage of volcanism.

As the volcanism temporarily ceased the region was subjected to faulting which was largely due to isostasy which reactivated the Citanduy Fault (see also the tectonic and basinal development section), with mainly dip–slip movement producing the eastern block sinking down relatively to the western block. This stage was then accompanied by the development of coral reefs (Kalipucang Limestone), restricted on the eastern flank of the Jampang outer arc (and other similar arc further to the east, northeast and also to the west). The growth of coral reefs is also accompanied by the deposition of the Penanjung and Pemali Formations by traction current in an open and quiet marine environment in the Middle Miocene time.

In the Upper Miocene, as the tectonism and volcanism were active again and subsequently followed by the availability of volcanic products to become the source of the sediments, and development of the sediment gravity flows which have been generated or initiated by earthquake activity as the agent for transporting and depositing the sediments derived from the volcanism above. These sediments (the lower part of the Halang Beds) were deposited largely in the basin to the east of the Citanduy Fault. The volcanic materials were mixed together with the sedimentary fragments derived from the older rocks and deposited in the arc trench gap. Volcanism waned out (or temporarily ceased) at the end of Middle Miocene, and sediment deposited in the basin was dominated by marl and calcarenites (upper part of the Halang Formation). In the Pliocene the volcanism was active again giving rise to development of thick volcanogenic sediments Kumbang Formation, which partly are deposited by sediment gravity flows, and partly by traction current, and the basin was much shallower in Middle Pliocene time. Further discussion on the tectonic and deposition development is chronologically given in the following section.

Depositional and Tectonic Development

Tectonic and depositional models proposed herein are simply based on the interpretation of data compiled from the previous geological works and the present study, including geological mapping, stratigraphy, geophysical investigation, structural and tectonic study, and other relevant geological studies. The present study stresses on the vertical sequences of sedimentary succession, provenance a
of the provenance of the sedimen
paleontology and direction of pal
depositing the sediments. These
lated and put together in an atten
tectonic models of the region.
depositional development will tl
in relation with tectonic develo
region.

Late Mesozoic

Tectonic evolution of the soutl
nesian archipelago is marked b
of the northwardly moving Ind
plate against the Asian Contin
which Sumatra, southern part
and West Java are included. S
sion of the collision is now ma
occurrence of melange complex
(Martojoyo and Suparka, 1977
Lok Ulo (Sukendar, 1974) in
and Pegunungan Meratus (Hami
which are thought to have bee
the outer swell on the ocean s
trench. The outer swell appear
been thrusted down the trench
Cretaceous time and is now bu
Eocene sediments (Fig. 9, 10). O
elements belonging to this tect
include, the volcanic chain whi
obscure, but it appears to be l
ten–Lampung region continuing
part of central of the southern
Kalimantan; and the Java Sea h
been formed as back deep basi
study area was located in the r
originally was a fore deep basin
Cretaceous time. At his stage
called the Bandung and Bogor Tr

Miocene

In Early Miocene, volcanism w
to be active in some part of th
pying originally by the Cretace
zone, and become the source a
sediments (Jampang Formation
part of this formation which c
limestone and spilitic fragments

are believed to have been derived from Jampang–Menoreh High. The volcanism was intermittently active during Lower Miocene time as indicated by the alternating of volcanogenic sediments and clastic rocks in the middle and upper part of the formation. During volcanism earthquakes were either intermittently or spontaneously active and generated sediment gravity flows to become the agent for transporting and depositing the Jampang Formation.

In the beginning of Middle Miocene, volcanism was temporarily ceased which was subsequently followed by folding and faulting older rocks. The Citanduy Fault has rectified in a dominant normal dip-slip movement, where the Jampang Block had been lifted and forming the so called Jampang-Menoreh Non-volcanic Outer Arc, relative to the sinking down eastern block. This is then subsequently followed by the development of coral reef particularly on the eastern flank of the Jampang Outer Arc, which is also accompanied by the deposition of the Pemali Formation and Penanjung Formation.in an open and quiet marine environment. These sediments are deposited by traction current. (Fig. 10).

In early Upper Miocene volcanism was intermittently active again producing andesitic volcanogenic sediments deposited abundantly in the basin to the east of the east of the Citanduy Fault. The earthquake activity accompanying the volcanism had generated and initiated the sediment gravity flows for transporting and depositing the sediments.

In late Upper Miocene volcanism appears to have waned out or temporarily ceased and the sediments deposited in the inter deep (arc–trench gap) has dominated by alternating of clastic rocks, marly tuff, marly mudstone, and calcarenite (upper part of the Halang Beds). These rocks are believed to have been derived from the older rocks further to the north.

In Lower Pliocene volcanism was highly active again and produced materials for the volcanogic sediments Kumbang Formation. Marl intercalations within the succession indicate that they have been deposited in marine environments, partly by sediment gravity flows and partly by traction current (Fig. 10).

As the volcanism ceased in Upper Pliocene, the region was subjected to folding and faulting episode. The whole region was spontaneously lifted, although some part of the area was a shallow marine basin in which the Tapak Formation was deposited.

In the Quaternary the subduction zone migrated further to the south (Fig. 13). The tertiary Java trench and the present active trench are both submarine trenches (Katili, 1973).

CONCLUSIONS

As most geologists have accepted the plate tectonic theory as fundamental geological knowledge, the present author has attempted to use the theory to the sedimentology and Tertiary tectonic development of the Pangandaran – Cilacap region.

The model proposed herein could therefore explain the development of basin from time to time, basinal filling, and sedimentary origin as well. The author believes that "the validity of any tectonic model depends on the how far and accurately the model could answer or explain or at least approach any geological problem arising the model".

1. The volcanogenic sediments of the lower Miocene Jampang Formation and the alternating coarse clastic and pelitic rocks of the Upper Miocene Halang Formation are deposited by sediment gravity flows.
2. The Jampang Formation have been deposited in an arc trench gap (interdeep basin), the lower part of the succession is believed to have been partly deposited in the upper slope of a foredeep basin. The succession appears to have been unconformably underlain by oceanic curst.
3. The sediment gravity flows were generated or initiated by earthquake shocks during and subsequently after volcanism. Thicker succession of volcanogenic sediments would imply that the volcanism was very active and sediment gravity flows have intensively

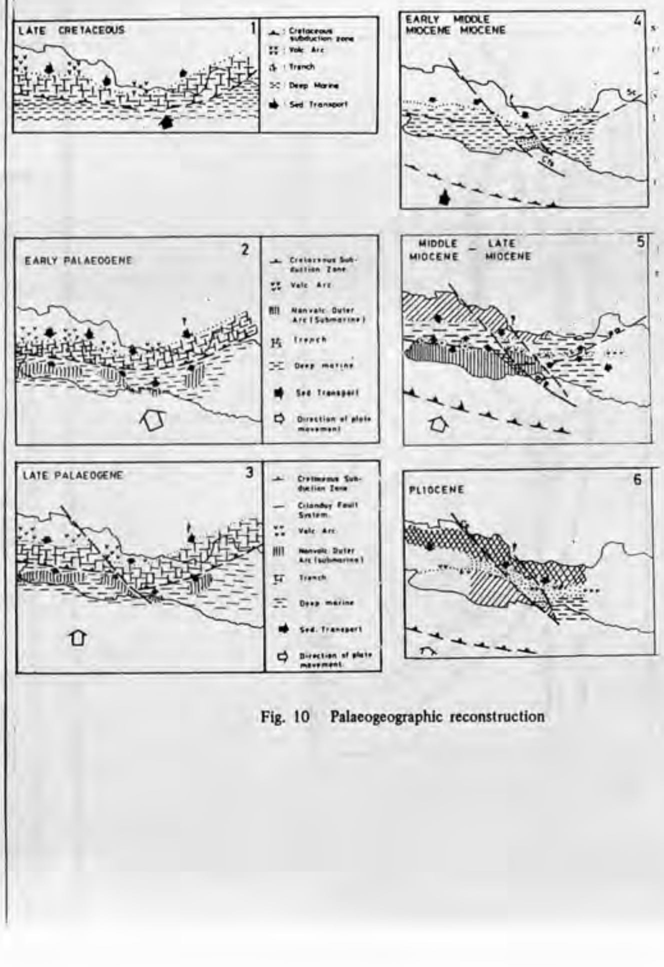


Fig. 10 Palaeogeographic reconstruction

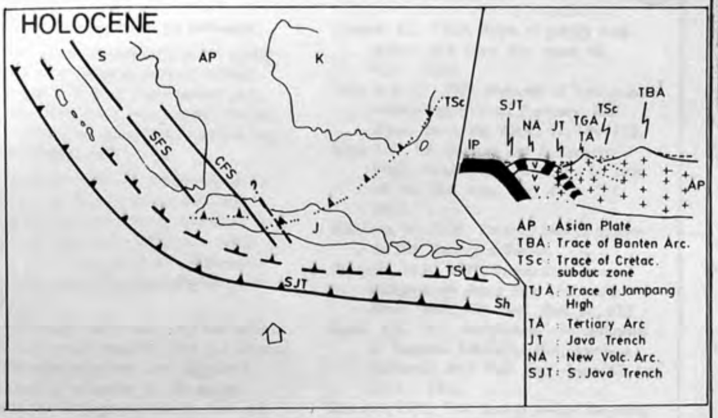


Fig. 11 Tectonic Configuration of South Western part of Indonesia at present time

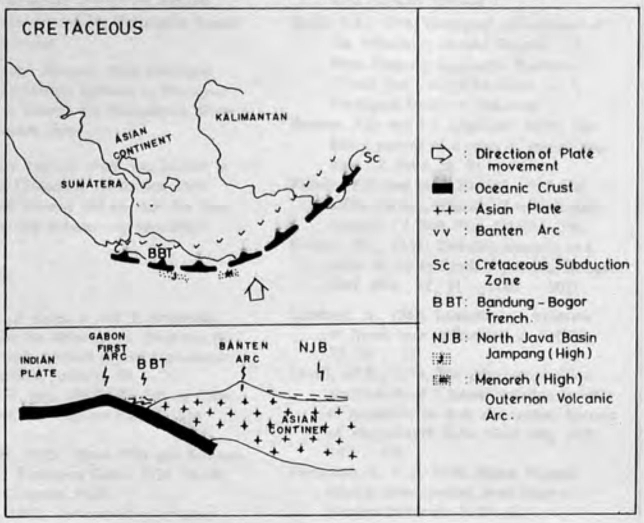


Fig. 12 Tectonic Configuration of South Western part of Indonesia Archipelago in Cretaceous time

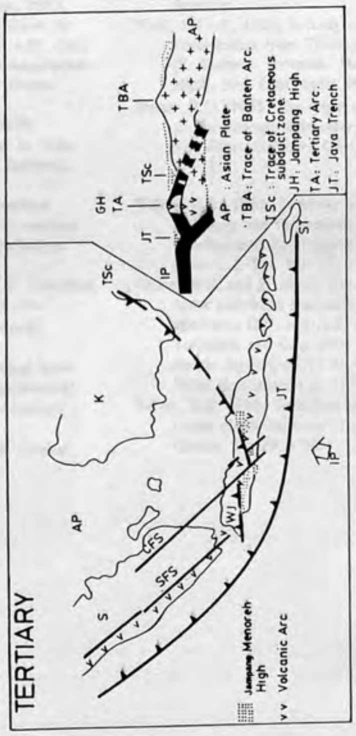


Fig. 13 Tectonic Configuration of South Western part of Indonesia in Tertiary

transported and deposited the sediments.

4. Volcanism was intermittently active during Neogene time, reaching peak of activity in the Lower Miocene and Pliocene times, and decreasing during early Upper Miocene time. Volcanism presumably ceased during Middle Miocene time.

5. The Citanduy Fault was presumably developed in Late Eocene with a dominant dextral strikeslip movement, reactivated in Middle Miocene with a dominant normal dipslip movement, and in Plio–Pleistocene apparently with a dominant reverse movement.

6. The Cretaceous subduction zone was initiated in a foredeep basin in which pre–Eocene and Paleogene sediments were deposited. The trench is indicated by the occurrence of the melange complex in Java and SE Kalimantan. The Bogor–Bandung Zones of Bemmelen (1949) are believed to have been part of the Cretaceous Trench (fore deep basin).

7. The Jampang Menoreh High developed after the Cretaceous collision in Paleogene time, and it became the Nonvolcanic Outer Arc in Miocene time.

8. The Tertiary volcanic chain was located in the belt of Cretaceous subduction zone in Java, and become the area for the Neogene and young sedimentary successions.

REFERENCES

Anketel, J.M., J. Celga, J. and S. Dzulynski, 1970, On the deformation structures in system with reversed deposity gradients: *Ann.Soc.Geol. Pologne*, 40, 3–30.

Bemmelen, R.W. van., 1949, *Geology of Indonesia*, vol. IA, Martinus Nijhoff, The Hague.

Bothe, A.Ch.D., 1929, Djiwo Hills and Southern Ranges : Excursion Guide, *IVth. Pacific Science Congress*, 1929.

Bouma, A.H., 1962, *Sedimentology of some Flysch Deposits:* Elsevier, Amsterdam: 168 pp.

Crowell, J.C., 1957; Origin of pebbly mudstones: *Bull. Geol. Soc.* Amer 68, 933 – 1010.

Dott, R.H. Jr., 1963, Dynamic of Subaquaeous Gravity Depositjonal Processes: *Bull. Amer. Assoc. Pet. Geol.*, 47, 104–128.

Flint R.F., J.E. Sanders, and J. Rongers., 1960, Diamictites, a substitute for symmictite: *Bull. Geol. Soc. Amer.* 701, 1089.

Hamilton, W., 1978, *Tectonic Map of Indonesia*, U.S. Geological Survey.

Hampton, M.A., 1972, Transport of ocean sediments by debris flow (Abstract): *Amer. Assoc. Pet. Geol. Bull.* 56, 622.

Haner, B.E., 1971, Morphology and sediments of Redondo Submarine Fan, Southern California: *Bull. Geol. Soc. Amer.* 82, 2413 – 2432

Kastowo, 1975, Peta Geologi lembar Majenang *Geol. Map of. Majenang Quad.* : Direktorat Geologi Bandung.

Katili, J.A., 1974, Geological environment of the Indonesian Mineral Deposits : A Plate Tectonic Approach: *Publikasi Teknik Seri Geologi Ekonomi*, no. 7, Geological Survey of Indonesia.

Kuenen, P.H. and C.I. Migliorini, 1950, Turbidity current as a cause of graded bedding : *J. Geol.* 58, 91 – 127

Kuenen, P.H. and H.W. Menard, 1952, Turbidity current, graded and non–graded deposits : J. Sed. Pet., 32, 83 – 96.

Kuenen, P.H., 1958, Turbidity currents as a major factor in flysch deposition: *Eclogae. Geol. Helv.*, 52, 51 – 1009 – 1021.

Lombard, A., 1963, Laminites : a structure of flysch type sediments : *J. Sed. Pet.* 33, 14 – 22.

Lovell, J.P.B., 1970, The palaeogeographical significance of 1 lateral variation in ratio of sandstone to shale and outher features of Aberystwyth Grits. *Geol. Mag.* 107: 147 – 158.

Martojoyo, S., et.al, 1978, Status Formasi Ciletuh dalam evolusi Jawa Barat : *Geologi Indonesia*, 5, 29–38.

Middleton, G.V., 1966, Experiments on density and turbidity currents. I Motion of the head : *Canad. J. Earth Sci.* 3, 523–546.

Middleton, G.V., 1967, Experiments on density and turbidity currents. III. Deposition of sediment: *Canad. J. Earth Sci.* 4 : 475 – 505.

Middleton, G.V. and M.A. Hampton, 1973, Mechanism of flow and deposition, *in* Middleton, G.V., and Bouma A.H., (eds) Turbidites and deep water sedimentation: *Pacific Section, SEPM Short Course Notes, Los Angeles*, 1 – 38.

Prakash, B. and G.V. Middleton, 1970, Downcurrent textural changes in Ordovician turbidite greywackes: *Sediment.* 14, 259 – 293.

Sanders, J.E., 1965, Primary sedimentary structures formed by turbidity currents and related sedimentation mechanism: *SEPM.* 12, 192 – 219.

Simanjuntak, T.O., 1977, Megacyclic Turbidites, *Unpubl. Thesis, University of New England*, Armidale, NSW Australia, 353 pp.

Simanjuntak, T.O., 1979, Peta Geologi Lembar Pangandaran, Jawa (in preparation): *Geological Research and Development Centre, Bandung.*

Sukamto, Rab., 1975, Peta Geologi Lembar Jampang dan Balekambang, Jaw Map of Jampang and Balekamb Quadr. : *Direktorat Geologi Ba*

Sukendar, A., 1974, Evolusi Geologi Tengah dan sekitarnya ditinjau Teori yang baru : *Unpubl. The Bandung.*

Vlerk, I.M.v.d., 1925, A study of the Foraminifera from Tidoengsche (E. Borneo): Wetensch. Meded. Mijnb. Ned. Oost. Indie, No. 3.

Walker, R.G., 1965, The origin and s of the internal sedimentary str turbidites: *Yorkshire Geol. Soc.* 1 – 31.

Walker, R.G., 1967, Turbidity sedim structures and their relationshi proximal and distal deposition ments : *J. Sed. Pet.* 37, 25 –

Walker, R.G. and E. Mutti, 1973, Tu facies and facies associations, *i* Middleton G.V. and A.H. Boun Turbidites and deep water sedi *Pacific Section, SEPM Short C* Notes, Los Angeles p. 119 – 1

Walker, R.G. 1976, Turbidites and a coarse clastic deposits: *Geoscie* Canada. 3 : 25 – 36.

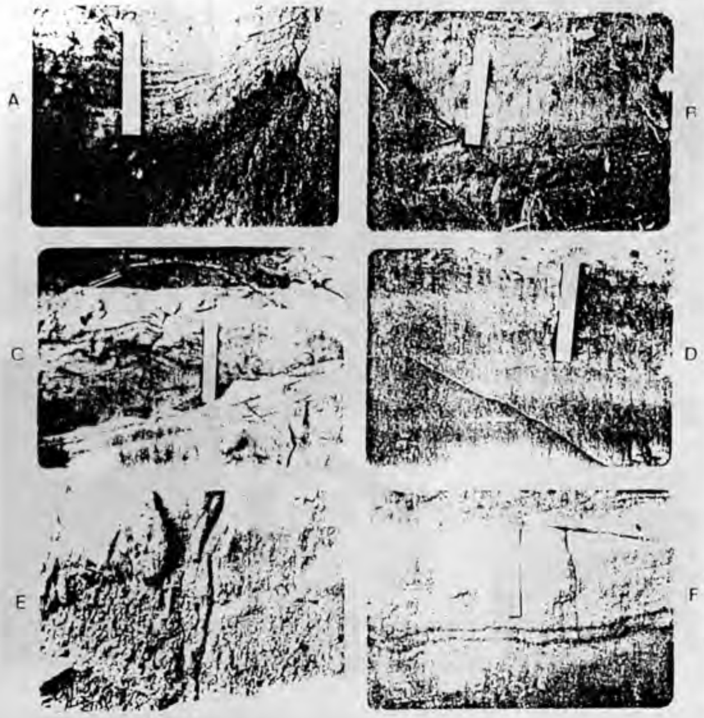


PLATE 1

A Pebbly arenite (Ta) and parallel lamination (Tb) forming a truncated sequence of Bouma in Jampang Formation.

B Parallel and current ripple lamination occurring in tuff beds of Nusakambangan Tuffs.

C Convolute lamination in Te intervals of turbidites in Nusakambangan Tuffs.

D Pebbly arenite with stratification in Jampang Formation.

E Submarine channel deposits in Jampang Formation. Note the disorganized conglomerate (bottom) with clasts predominantly of volcanics and subsidiary limestones and silty pelitic and the overlying of massive and structureless pebbly lithic arenite.

F Well graded pebbly tuff arenite (Ta) and parallel laminated tuff arenite (Tb) intervals of turbidite in Nusakambangan Tuffs.

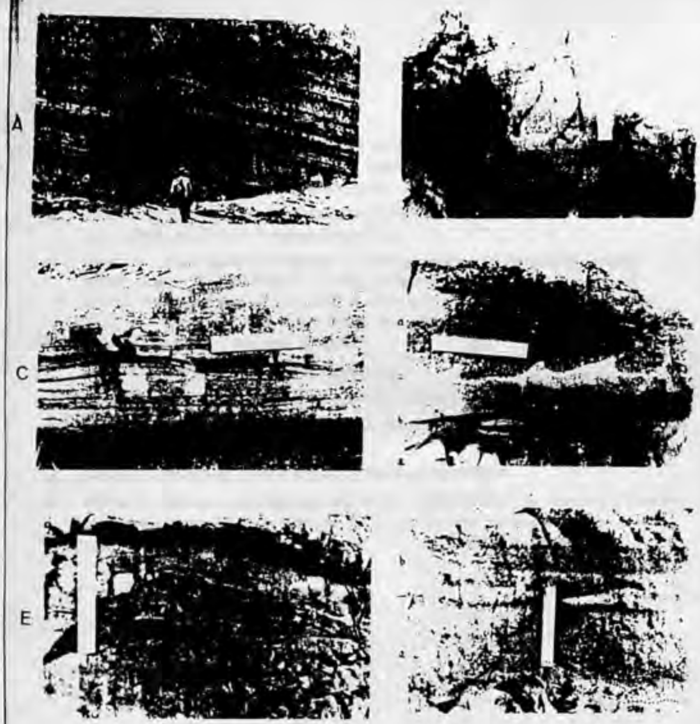


PLATE 2

A Parallel sided beds of tuffaceous arenites in the upper part of Jampang Formatio

B Truncated sequence forming by graded arenite (Ta) and parallel laminated (Tb) i Jampang Formation.

C Base–cut–out and truncated sequence is composed of parallel laminated arenite (Tb) and thinly current ripple laminated layer (Te) in Jampang Formation.

D Load casting or plume structure in Jampang Formation.

E Diffuse stratification in very coarse lithic arenite of Jampang Formation.

F Truncated sequence in Jampang Formation. Note the erosional surface at base of the Ta interval.

PLATE 3

A Steeply dipping succession of organised conglomerate (OC), diamictite (d), para-conglomerate (pc) and arenite (a) with thinly bedded silty tuffs (s) in Jampang Formation. The rocks are all volcanogenic origin.

B Moderately sorted organised conglomerate with clasts dominated by volcanics and minor lithic arenites in Jampang Formation.

C Development of truncated sequence of Bouma forming by moderately graded pebbly arenite (Ta) on bottom, parallel laminated arenite (Tb) at middle and current ripple laminated fine graded arenite on top (Tc), in Jampang Formation. Note the presence of mud flakes within the upper part of the Ta, and the erosional surface on the base of Ta.

D Lensoidal basaltic lava in Jampang Formation.

E Organised conglomerate similar to B in the Jampang Formation.

F Volcanic breccia with clasts of pyroxene andesite in the upper part of the Jampang Formation.

G Massive, structureless pebbly arenite in Jampang Formation.

H Stratified organised conglomerate with weakly oriented clasts in Jampang Formation. Note the presence of layer clasts at the bottom of the stratification.

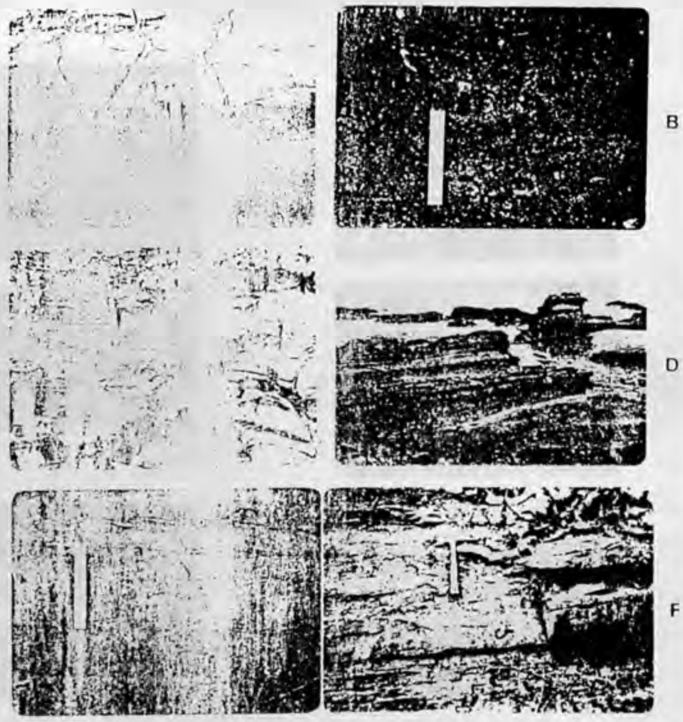


PLATE 4

A Convolute lamination of Tc interval in Nusakambangan Tuffs.

B Development of complete (?) sequence in Jampang Formation. The sequence is graded from pebbly arenite (Ta) on the bottom to coarse grained and parallel laminated lithic arenite (Tb) through current ripple laminated fine arenite (Tc) to parallel laminated silty pelitic rocks (Td) on top. Te is possibly present but it is difficult to distinguish from Td intervals.

C Slightly folded thinly laminated tuffs in Nusakambangan Tuffs.

D Parallel sided beds of tuffs, tuff arenites in Nusakambangan Tuffs.

E Development of Ta–d sequence (complete ?), similar to B in Nusakambangan Tuffs.

F Full apart structure in Jampang Formation.

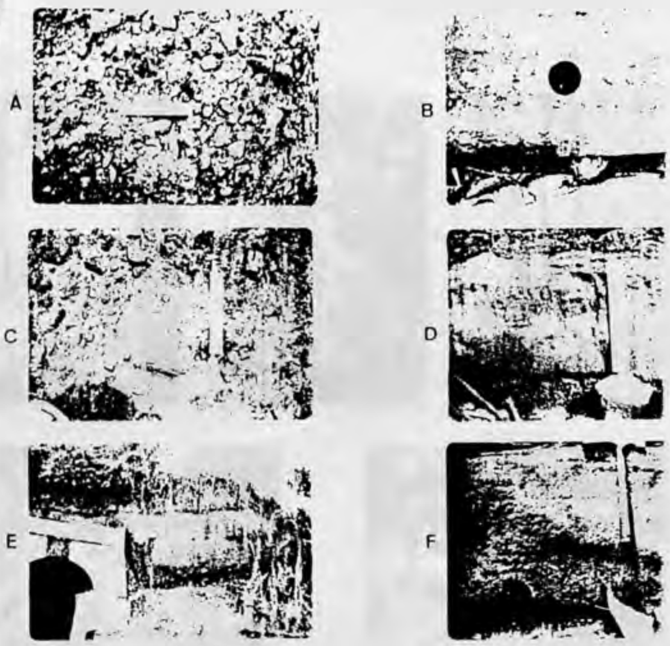


PLATE 5

A Moderately sorted organised conglomerate in Jampang Formation. The cl is dominated by volcanic rocks.

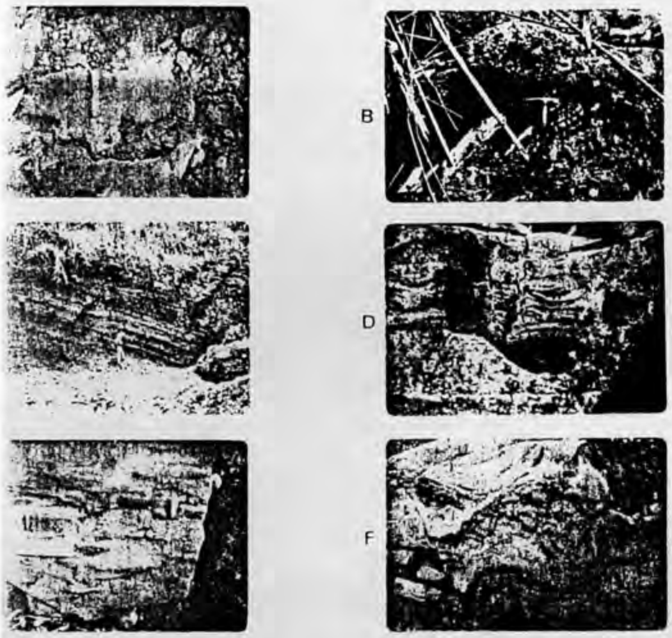
B Pebbly arenite shows stratification and grading in Jampang Formation.

C Disorganised conglomerate with clasts consisting almost entirely of pyrox andesite in Jampang Formation.

D Development of Tab in arenite bed of Jampang Formation.

E Graded and stratified volcanogenic clastic rocks in Jampang Formation.

F Massive, and stratified pebbly–lithic–arenite in Halang Formation. Note presence of erosional surface at the base of bed.



LATE 6

A Diamictite in Jampang Formation.

B Diamictite in Jampang Formation. Note the presesence of large mud flakes.

C Stratified volcanogenic rudite in Jampang Formation.

D Convoluted current ripple lamination in Tc interval in Halang Formation.

E Weakly developing flute cast on the base of arenite beds in Halang Formation.

F Deformed convolute lamination in Tc interval in Halang Formation. It appears that there is a gradational deformation of the Tc intervals from D to F.

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Tectonic Relationship Between Geologic Provinces of Western Sulawesi, Eastern Sulawesi and Banggai–Sula in the Light of Sedimentological Aspects

Rab Sukamto & T.O. Simandjuntak

ABSTRACT

The geological complexity of Sulawesi Island and its surroundings has intrigued earth scientists for a long time. In attempts to explain the tectonics of the region many geologists have advanced various theories and hypotheses. Recent systematic geological mapping and other related research on the island have provided a better understanding of the tectonic development of the island. On the basis of sedimentological aspects, the tectonic development of Sulawesi Island and its surroundings can be explained.

Late Cretaceous to Eocene flysch-type sediments derived from volcanic island arcs were deposited in Western Sulawesi Province and Jurassic to Cretaceous pelagic sediments were deposited in Eastern Sulawesi Province in association with ophiolite, while Triassic to Cretaceous continental-derived sediments were deposited on a stable shelf of the Banggai-Sula microcontinent which has a basement of Carboniferous metamorphics and Permo-Triassic plutonic rocks.

During the Cretaceous the flysch of the Western Sulawesi Province was deposited in a forearc basin, the pelagic sediments of the Eastern Sulawesi Province was deposited in a deep oceanic environment and the clastics of the Banggai–Sula Province was deposited in a continental shelf environment. Those geological elements are all closely related within the plate tectonic framework of the region.

INTRODUCTION

This paper discusses tectonic evolution of Sulawesi Island and its surroundings in the light of sedimentological aspects. The discussion is based upon the available data, particularly on new data collected by the geological mapping project since 1971.

Most earth scientists believe that the geological complexity of the Indonesian island arc system is a result of interaction of three major crustal elements, the northward-moving Indian Ocean Plate, the westward-moving Pacific Plate and the south-southeastward-moving Eurasian Plate. Sulawesi Island is situated in the central part of the Indonesian archipelago (Fig. 1) and is one of the most complicated regions from the plate tectonic point of view.

Many investigations on the geology and mineral resources of Sulawesi Island and its surroundings were carried out between 1887 and 1968. On the basis of these data the major tectonic features of Sulawesi Island and its surroundings were described in very different ways by Ahlburg (1913), Abendanon (1915, 1917, 1918), and Brouwer (1930), and geological synopses of Sulawesi have been advanced by Rutten (1927), Brouwer (1934, 1947), Kundig (1956), van Bemmelen (1949) and Visser and Hermes (1962). Based on plate tectonic theory the tectonic evolution of the Indonesian Archipelago of eastern part of Indonesia has been described by Hamilton (1970, 1973, 1979), Katili (1970, 1971, 1972, 1974, 1978), Audley-Charles *et al.* (1972), Gribi (1973) and Sukamto (1975a, 1975c). Carey

*) Paper originally presented to the IV[th] Geosea Conference in Manila, November 1981

(1975) unravelled the complexity of the structural pattern of the Indonesian archipelago by the reversal of three processes, viz. Tethyan torsian, the Ninetyeast and peri-Pacific dextral megashear, and a general dispersion.

Since 1971 a systematic geological mapping project run by the Geological Survey of Indonesia (GSI) in the framework of the First Five–Year Development Programme, and the influx of foreign investment in the fields of mineral and energy resources, have rapidly produced new data giving rise to a better understanding of the geological development of the island.

Sukamto (1975a, 1975c) briefly discussed the structure of Sulawesi in the light of plate tectonic theory and concluded that: 1) The Banggai–Sula Province represents a continental plate that had become stable since Middle Mesozoic times, and was located close to the Eastern Sulawesi Province by the Triassic, 2) The huge masses of ophiolite in the Eastern Sulawesi Province suggest to have occurred during Triassic times, which was then subducted westward in Late Cretaceous – Early Tertiary time producing glaucophane schist in the western part of Eastern Sulawesi Province, and Early Tertiary volcanics of alkaline to calc-alkaline in composition associated with granitic intrusives within the Western Sulawesi Province, 3) In the Western Sulawesi Province a west-dipping subduction had become inactive since the Middle Miocene in the southern part, while in the northern part a north-west-dipping subduction is still active until the present.

Simandjuntak (1980) described the tectonic Wasuponda Melange in the central part of Sulawesi, and suggested that this melange wedge is a surface expression of a Cretaceous west-dipping subduction zone. In a later paper (1981) he also discussed sedimentological aspects of particularly Mesozoic strata in the eastern part of Sulawesi, concluding that a considerable thicknesses of Mesozoic continental-derived sediments deposited on the continental shelf of the Banggai–Sula ultramafic belt in Eastern Sulawesi. Those sediments were considered to be prospective for hydrocarbons.

The present authors on the basis of field data recorded during the geological mapping project in various parts of Sulawesi and its surroundings

Fig. 1 Location of Sulawesi island and its surroundings in the Indonesian Archipelago

attempt to relate the tectonic evolution of the region with the sedimentary processes since late Mesozoic to Neogene times within each province. It will be shown that there are very distinctive sedimentary successions deposited in each province, particularly during the Cretaceous viz. thick flysch-type sediments in Western Sulawesi Province, a pelagic deposit in Eastern Sulawesi Province and a very thick continental shelf deposit in Banggai–Sula Province. Paleogene sedimentary succession in each province are also distinctive.

Dr. G.P. Robinson and D. Trail has critically read the manuscript and have given valuable advice, for which the authors are duly grateful.

GEOLOGICAL SETTING

On the basis of lithological association and structure of the region, Sukamto (1975a, 1975c) divided Sulawesi Island and its surroundings into three geological provinces: Western Sulawesi Province, Eastern Sulawesi Province and Banggai–Sula Province (Fig. 2). In terms of orogeny the eastern province is relatively older than the western one.

The Western Sulawesi Province is characterized by Tertiary plutono-volcanic rocks in association with Tertiary and Quaternary sediments (Sukamto, 1975a, 1975c). The plutonics are granitic-dioritic rocks of Late Miocene to Pleistocene age (Fig. 3A); the volcanics are largely calcalkaline and minor alkaline rocks ranging in age from Paleocene to Pleistocene, though volcanoes are still active in the northern part of the province (Fig. 3B+C). Marine sediments and volcaniclastics were deposited intermittently during Paleocene to Holocene time. In the southern part the Tertiary rocks are underlain by a thick sequence of Late Cretaceous flysch type sediments (Fig. 3B). The flysch type sediment, over 200 m in thickness, are in turn underlain by a melange complex (Sukamto, 1981). The flysch was deposited continuously possibly from Late Cretaceous to Eocene time in the northern part of the province, and probably represents a sedimentary sequence deposited in a forearc basin.

The Eastern Sulawesi Province is composed of ophiolites associated with Mesozoic pelagic sediment and melange in the eastern part and metamorphic rocks in the western part (Fig. 3A+B). The ophiolites consist largely of dunite, harzburgite, lherzolite, serpentinite, wehrlite and minor gabbro, diabase, basalt and diorite (Suriaatmadja et al., 1972). The sequence is better developed in the north; in the central and south the ophiolites suite is mostly incomplete or disrupted (Simandjuntak, 1981). The pelagic sediments consist of alternating carbonate, radiolarian chert and red shale which were deposited in the central part consists of blocks of ophiolite, pelagic sediment and metamorphics, in a matrix of red scaly clay (Simandjuntak, 1980). The metamorphics in the west consist of a variety of schists, variously in the amphibolite-epidote, glaucophane-lawsonite or greenschist facies (de Roever, 1947).

The Banggai–Sula Province is characterized by a basement complex of Carboniferous metamorphics and Permo-Triassic plutonic rocks (Fig. 3A), overlain by a Mesozoic continental derived sedimentary succession containing *ammonites, belemnites,* and *pelecypods* (Sukamto, 1974). A sequence of coarse clastics of possibly Late Triassic age is

Fig. 2 Map showing boundaries of geologic and administrative provinces

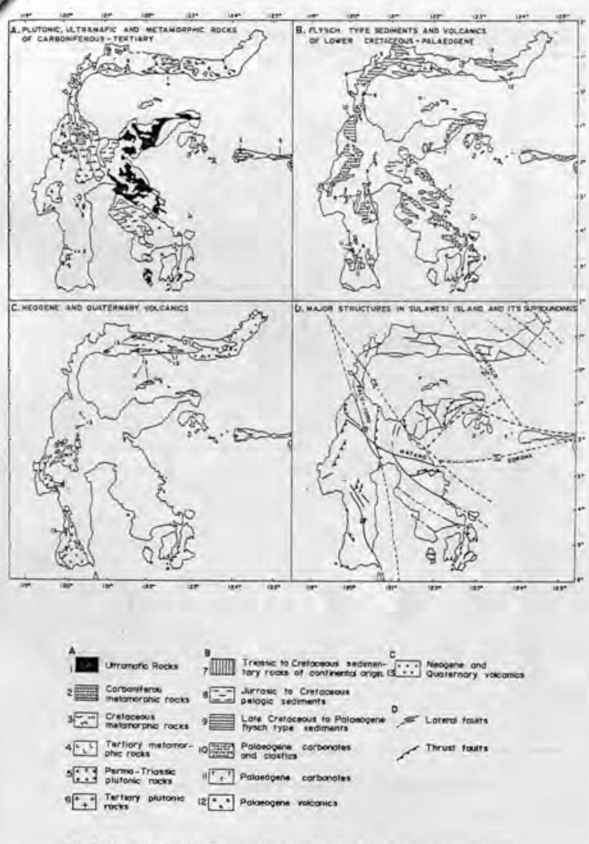

Fig. 3 Maps showing distribution of rocks in Sulawesi Island and its surroundings

conformably overlain by Jurassic fine clastics and Cretaceous calcareous rocks. Detritus of granite from this province extends into the east of the central part of the Eastern Sulawesi Province.

The tectonic pattern of Sulawesi and its surroundings (Fig. 3D) has been described briefly by Sukamto (1975a, 1975c). Evidence for a strong Middle Miocene orogenic phase can be observed in the Sulawesi region. In addition, a Late Pliocene orogenic phase is also evident, and presently active tectonism can also be observed. Orogenies prior to the Middle Miocene orogeny occurred locally during Early Miocene, Oligocene and Early Eocene; and occurred on a regional scale during the Triassic in the Banggai–Sula Province.

Intense folding succeeded by overthrusting took place during the Middle Miocene in the East Arm and in the central part of Western Sulawesi Province. The present regional tectonic pattern of Sulawesi is dominated by strikeslip faults and overthrusts. The Palu–Koro fault is a sinistral strikeslip fault, extending more than 750 km (Tjia, 1973). The sense of movement of this fault corresponds with those of the Matano and Sorong fault zones (Fig. 3D). On the other hand the Gorontalo fault zone is dextral (Katili, 1970). The overthrust pattern is mostly consistent with the trend of the Banggai–Sula Province. The direction of movement of the strikeslip faults and overthrusts indicate that the westward push of the Banggai–Sula Province was responsible for the lateral compression.

DEVELOPMENT OF SEDIMENTATION AND VOLCANISM

Western Sulawesi Province

The very thick sequence of Cretaceous to Eocene flysch in the Western Sulawesi Province (Fig. 3B, Fig. 4) have been mapped as the Marada Formation (van Leeuwen, 1979; Sukamto & Supriatna, 1981), Balangbaru Formation (Sukamto, 1981), Latimojong Formation (Djuri & Sudjatmiko, 1975) and Tinombo Formation (Brouwer, 1947; Sukamto, 1975b; Ratman, 1976; Trail, 1974; Apandi, 1974). The flysch lies unconformably on a metamorphic complex basement in the central and northern part while in the southern part they are underlain by a melange complex with metamorphic and ultramafics components.

The Marada and Balangbaru Formations were deposited in the south and the Latimojong Formation in central part of the province during the Late Cretaceous. These formations are characterized by alternating arenites and pelitic sediments. Well-bedded arenites of a few centimeters to tens of centimeters in thickness alternate with well-laminated shale. Turbidite features are commonly observed in these rocks.

Arenites and shale of the Marada Formation are intercalated by calcarenite, tuff, lava, volcanic breccia and conglomerates. The clasts in the conglomerates are basalt and andesite. The arenites range from arkosic to wackes. Lava and volcanic breccia are highly propylitized. Fossils of *Globotruncana* from calcarenite indicate an Upper Cretaceous age and deep neritic environment (van Leeuwen, 1979). The formation is more than 100 m thick.

Alternating arenites and shale in the Balangbaru Formation occurs mainly in the middle part of the sequence. The rocks in the lower part of the succession contain intercalations of polymict breccia and conglomerates while the upper part contain polymict conglomerates. The arenites range from arkosic to wackes with layers of a few cm to tens of cm in thickness. Intercalation of basalt occurs in the middle and lower parts. The components of the conglomerates are basalt, andesite, diorite, mudstone, silicified tuff, schist and quartz. Components of breccias and conglomerates in the lower part are schist, diorite, granite, quartz, serpentinite, arenites, wackes, limestones, and chert. *Globotruncana* identified from shale indicates an Upper Cretaceous age and neritic environment. The unit is about 2000 m in thickness, and is intruded by dykes, sill and stocks of dioritic, andesitic, and basaltic rocks.

Arenites and shale of the Latimojong Formation contain intercalations of lenses of basaltic to andesitic lavas and conglomerates. Some rocks of this formation have been metamorphosed into slate, phyllite and quartzite. Fossils of Cretaceous age identified from boulders have been suggested by Brouwer (1934) to be derived from this formation. Apparently the depositional environment of these

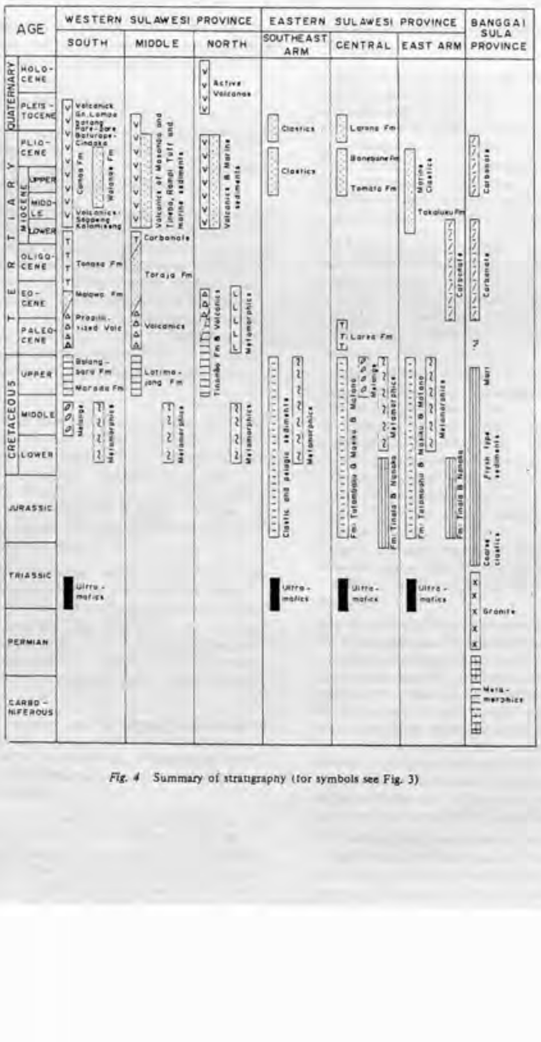

Fig. 4 Summary of stratigraphy (for symbols see Fig. 3)

rocks is closely related or similar to the Balangbaru and Maranda Formations. The formation attains a thickness of more than 1000 m and is intruded by dykes and stocks of basaltic to granitic rocks.

The flysch-type sediments developed in the northern part of Western Sulawesi Province (Fig. 3B) were deposited possibly during Cretaceous to Eocene times (Sukamto, 1975, 1975c) and are called the Tinombo Formation (Fig. 4) by Ahlburg (1913) and Brouwer (1934). The formation consists of alternating arenites and shale, gray, black and red on color with intercalations of conglomerate, limestone and tuffaceous arenites. Some arenites contain detritus of metamorphic rocks. Some of the rocks have been metamorphosed into slate, phyllite and quartzite. The limestone intercalation in the upper part of the succession contains fossils of *Nummulites* which indicated an Eocene age and shallow open shelf environment. Marine volcanic rocks of basaltic, spilitic to andesitic composition consisting of breccia, tuff and pillow lavas occur in some parts of the formation.

Volcanics of Palaeocene age occur in restricted areas in the south and have been mapped as propylitized volcanics (Fig. 4) in the Bantimala region (Sukamto, 1981) and Langi Volcanics in the Biru area (van Leeuwen, 1979). The volcanics consist of lavas (partly pillows), breccias and tuffs of basaltic to andesitic and trachytic composition with intercalations of arenites, shale and limestone. The limestone and pillow lavas indicates a submarine environment.

During Early Eocene times a sequence of terrestrial deposits (Malawa Formation, Sukamto, 1981) was deposited in the south and shallow marine deposits developed in the central part (Toraja Formation, Djuri & Sudjatmiko, 1975). The Malawa Formation consists mainly of arenites and shale with coal intercalations. The Toraja Formation is characterized by red shale alternating with quartzose arenite which is intercalated by nummulitic limestone and conglomerates. The sediments were deposited continuously from the Eocene to Oligocene (Fig. 4).

A thick carbonate sequence was deposited over a large area in the south and central part during the Late Eocene to Early Miocene. The carbonate sequence in the southern part was deposited continuously from Late Eocene to Early Miocene and has been mapped as the Tonasa Formation, while in the central part it developed only during the Early Miocene.

Volcanic rocks of Early Miocene age occur in some places along the Western Sulawesi Province. The rocks consist of breccias, lavas, and tuff with basaltic to andesitic composition. The volcanics were apparently deposited in a subaqueous environment as indicated by the occurrence of pillow basalt.

A very thick sequence of volcanic rocks deposited during Middle Miocene to Pliocene times extends over large area along the Western Sulawesi Province (Fig. 3C, Fig. 4). The volcanics consist of breccia, lava and tuff and are associated with marine sediments such as calcareous volcanoclastics, shale, marl and limestone. The composition of the volcanic rocks is basaltic to andesitic in the lower part (Middle Miocene), alkali basalt in the middle part (Upper Miocene), and andesitic to trachytic in the upper part (Pliocene). This rock association has been mapped in the south as the Camba Formation (4000–5000 m) by Sukamto (1981), in the central as the Masamba and Tineba Volcanics and Rampi Tuff by Simandjuntak *et al.* (1981), and in the north as Neogene Volcanics by some workers (Apandi, 1974; Effendi, 1974; Trail *et al.* (1974; Ratman, 1976). The volcanics were deposited in submarine to subaerial environments.

Formation of Kuroko type copper deposit in this volcanics is indicated in the Sangkaropi region, Central Sulawesi (Sunarya & Yudawinata, 1980).

During deposition of the Neogene volcanics, in parts of the Western Sulawesi Province graben-like basins developed in which the clastic and carbonate rocks were deposited. The sediments consist partly of molase type deposits; i.e. coarse conglomerate, marine tuff, arenites, shale, marl and coralline limestone (Fig. 4).

Subaerial volcanic rocks of possibly Pleistocene age occur in a large area at the southern end of the Western Sulawesi Province, and some smaller areas in the central part. The composition of the volcanics is basaltic, andesitic and trachytic. The Quaternary volcanic rocks extend over large areas in the northern part of the province (Fig. 3C). In this area some of volcanoes are still active (Fig. 4).

The sedimentation and volcanism described earlier indicate that the Western Sulawesi Province represents a Cretaceous forearc basin which developed into a volcanic arc since the Early Tertiary.

Eastern Sulawesi Province

A Late Cretaceous deep-sea sedimentary sequence developed in the Eastern Sulawesi Province (Fig. 3B) synchronously with the flysch-type deposits in the Western Sulawesi Province (Fig. 4).

The Late Cretaceous deep-sea sedimentary rocks in the Eastern Sulawesi Province are characterized by alternating calcilutite and radiolarian chert in the lower part and a greater amount of calcilutite in the upper part. These deep-sea sediments have been mapped as the Upper Matano Formation by Brouwer (1947) and the Matano Formation by Simandjuntak et al (1981). The calcilutite contains fossils of *Globotruncana* and *Heterohelix* of Late Cretaceous age. It attains a thickness of about 500 m.

The Cretaceous sedimentary sequence is conformably underlain by the Masiku Formation (Simandjuntak et al., 1981) or Lower Matano Formation of Brouwer (1947). The Masiku Formation consists of calcilutite, interbedded with radiolarian chert and intercalations of wackes and shale. As well as bedded chert there are also nodules within the calcilutite beds.

The Cretaceous rocks in the central part are conformably underlain by a succession of bedded calcilutite and chert, with intercalations of calcareous shale and lithic arenites (Fig. 4). The unit has been mapped as the Tetambahu Formation (Simandjuntak *et al.* (1981). It attains a thickness of 500 m, and is thought to be of Jurassic age.

In the southern part the Jurassic rocks consist of shale, lithic arenite, calcilutite and radiolarian chert with sparse molluscs. Some of the rocks are slightly metamorphosed into slate, phyllite and quartzite.

The Jurassic–Cretaceous sequence has been strongly deformed and faulted and hence its original thickness is not known, but the authors suggest a thickness of at least some hundreds of meters. Its contact with the basement is often faulted. The authors believe that these formations were deposited on top of the ultramafic basement.

A sequence of sedimentary rocks predominantly consisting of bedded limestone intercalated by calcarenite conformably overlies the Matano Formation. The sequence is mapped as the Lerea Formation (Simandjuntak *et al.*, 1981) in the central part of the Eastern Sulawesi Province, and dated as a Paleogene age and deposited in a relatively shallow open marine environment. It attains a thickness of about 150 m.

The Lower Miocene sediments consisting of bedded limestone and calcarenite with marl intercalations is mapped as the Takaluku Formation in the East Arm (Simandjuntak *et al.*, 1981). The fossils of foraminifera indicate a relatively shallow marine environment. It attains a thick-thickness of at least 200 m.

The upper Miocene sediments are mapped as the Tomata and Bone-bone Formations (Simandjuntak *et al., op. cit*) in the central part, and in the south as a "Molasse deposit" (Sukamto, 1975b, Kartaadipoetra & Sudiro, 1973). The rocks consist mainly of arenites, conglomerate and subsidiary shale and marl. The Bone-bone Formation contains more abundant coarse clastics, while the Tomata Formation contains finer clastic. The upper part of Tomata Formation contains lenses of lignite and plant remains which indicate a terrestrial origin. The Tomata and Bonebone Formations were deposited up to the Middle Pliocene.

Late Pliocene to Pleistocene sedimentary rocks in the Eastern Sulawesi Province are characterized by a terrestrial environment of deposition and were deposited in some isolated areas such as Larona and Pindolo in the central part (Simandjuntak *et al., op. cit*), as well as the Wawotobi and Wanonumbotea areas in the south (Sukamto, 1975b). The rocks consist of coarse to fine detrital materials derived from the surrounding older rocks. The rocks in these isolated areas possibly were deposited in graben-like basins. In some areas similar clastics may have been deposited in a coastal environment.

The sedimentation mentioned above indicates that during Jurassic to Cretaceous the Eastern Sulawesi Province was a deep-sea basin, which was shoaling since the Paleocene and become a terrestrial environment at least from the Late Pliocene.

Banggai–Sula Province

The Tinala and Nanaka Formations (Fig. 3B; Fig. 4) might belong to the Banggai–Sula Province (Simandjuntak, 1981). The Tinala Formation consists of limestone, shale and lithic arenite intercalations. The limestone contains fossils of *Halobia* and *Ammonites*, which indicate Triassic age and a continental shelf environment. The Nanaka Formation is characterised by the occurrence of quartzose arenite and conglomerate containing clasts of pink granite which are thought to have derived from the Banggai–Sula basement complex (Sukamto, 1975a, 1975c).

These two formations are quite distinct from the other Mesozoic sediments in the Eastern Sulawesi Province, however are closely similar to the Mesozoic sediments of the Banggai–Sula Province (Fig. 4). In Banggai–Sula Province a sequence of coarse clastic (sandstone and conglomerate) derived from metamorphics, granitic and volcanic rocks of Carboniferous–Triassic basement complex (Sukamto, 1974) is widely distributed (Fig. 4). Arkosic sandstone and conglomerate which have among their constituents fragments of pink granite, extend as for as Toeli on the East Arm of Sulawesi (Sukamto, 1975a, 1975c), and were later includes within the Tinala and Nanaka Formations (Simandjuntak *et al.*, 1981).

Again in Banggai–Sula a sequence of fine clastics or flysch type sediments consisting of shale with intercalated sandstone layers conformably overlie layers of sandstone and conglomerate mentioned earlier. This sequence abounds in *Amonite, Belemnite, Pelecypod* and cone-in-cone concretions, and Late Jurassic age (Sukamto, 1974; 1975a, 1975c). These layers change gradually upward into an overlying sequence of calcareous sandstone, marl and shale containing a few molluscs fossils. The age of this sequence is Cretaceous

Limestone and marl layers alternating with sandstone and conglomerate of pre-Middle Miocene age were deposited in the southern part of the East Arm of Sulawesi and Peleng Island (Fig. 2; Fig. 3B; Fig. 4). Sedimentation on Peleng Island started in Miocene time, whereas in the East Arm it began in the Eocene. Coral reefs of Late Miocene to Pleistocene age deposited post-Middle Miocene orogenesis are widespread on the Banggai–Sula Province (Sukamto, 1974, 1975a, 1975c).

The sedimentation that occurred in the Banggai–Sula Province from Triassic to Cretaceous indicates that the area was a continental shelf environment, which was uplifted but locally remained as a shallow sea since the Early Tertiary.

TECTONIC DEVELOPMENT

Late Cretaceous

The tectonic development of the Western and Eastern Sulawesi Provinces is closely related to the tectonic development of Banggai–Sula Province. During the Late Cretaceous a thick sequence of flysch-type sediments were deposited in broad areas along the Western Sulawesi Province (Fig. 5A). These flysch-type sediments are unconformably underlain by the melange complex in the south part and by metamorphic complex basement in the central and the north parts. The sediments are commonly associated with lavas and pyroclastics indicating that this rock association was derived from volcanic island arcs and deposited in a forearc basin area.

At the same time the region of the Eastern Sulawesi Province developed as a deep-sea basin, in which pelagic sediments were deposited since Jurassica time on the ophiolite basement. It is very possible that the Cretaceous deep-sea basin in the Eastern Sulawesi Province was separated by a trench from the Western Sulawesi Province. The trench was possibly a surface appearance of a westerly-dipping subduction zone, in which the Wasuponda Melange accumulated (Simandjuntak, 1980). The subduction initiated magmatism along the western Sulawesi Province. The metamorphic rocks occurring along the western part of the Eastern Sulawesi Province are believed to have formed during this Cretaceous subduction.

By contrast the Banggai–Sula Province was part of a continental shelf since the Early Mesozoic, in which Late Triassic to Cretaceous clastics were deposited. The core or basement of the continent consists of Carboniferous metamorphics and Permo-Triassic plutonics.

Paleogene

Development of flysch-type sediments in the Western Sulawesi Province terminated in the southern part, while in the northern part con-

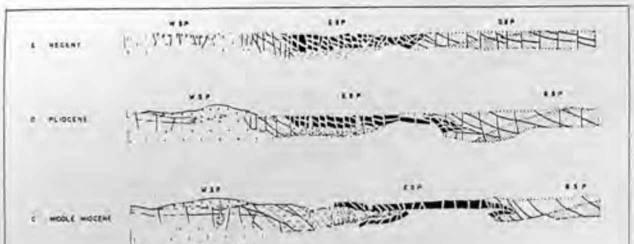

Fig. 5 Palinspatic sections across Central Sulawesi and Banggai–Sula Islands showing tectonic development of the region related to sedimentological aspects

tinued until Eocene times (Tinombo Formation, Sukamto, 1975a, 1975c). Volcanoes were locally active during the Paleocene in the south and during the Eocene in the Central and north (Fig. 5B). Deposition of a thick carbonate rocks (Tonasa Formation) occurred in a large area in the south during the Eocene to Miocene indicating that this part of the area was a stable shelf.

Since the Palaeocene the Eastern Sulawesi Province had appeared to be shoaling and shallow-water carbonates had been deposited in this environment (Lerea Formation, Simandjuntak, 1981). Deposition of carbonate rocks in this area continued up to the Lower Miocene (Takaluku Formation).

In the western part of the Banggai–Sula Province a thick sequence of carbonates intercalating with clastics were deposited in large areas. These carbonates were deposited until the Middle Miocene.

The westerly-dipping subduction zone which commenced from the Cretaceous produced the Early Tertiary volcanics in the Western Sulawesi Province and a shoaling process of the sea in the Eastern Sulawesi Province as well as Banggai–Sula Province (Fig. 5B).

Neogene

The wide distribution of volcanic products indicate that strong volcanism recurred since the Middle Miocene in the Western Sulawesi Province (Fig. 5C). The volcanic rocks were initially deposited in a submarine environment and then locally became terrestrial in the Pliocene. The volcanism terminated in the Lower Quaternary in the south but continued until recent times in the northern part of the province.

Strong magmatism in the Western Sulawesi Province during the Middle Miocene was apparently coincident with the squeezing process of the rocks within the Eastern Sulawesi Province (Fig. 5C) due to the westward movement of the Banggai–Sula microcontinent. This tectonic episode has uplifted and thrusted most of the material within the Eastern Sulawesi Province. The metamorphic rocks were thrusted westward onto the Western Sulawesi Province; likewise the ophiolite rocks were also thrusted and imbricated with associated rocks possibly including the melange, but to an opposite direction, e.g. eastwards into the Mesozoic and Paleogene sediments of the Banggai–Sula Province.

During the uplifting of the whole region of Sulawesi which commenced from the Middle Miocene, block-faulting was initiated in various places to form graben-like basins (Fig. 5D). In Pliocene time the whole region was subjected to block faulting and the major fault, such as the movement afterward initiated the present morphology of Sulawesi Island (Fig. 5E). This tectonic event produced a shallow and narrow marine basin in some parts of the region and some isolated basins inland. Coarse clastic rocks were deposited in these basins and formed the so-called Sulawesi Molasse.

The Middle Miocene tectonic event also bent the Western Sulawesi Province into its present curved form and exposed the metamorphics within the neck of the island.

REFERENCES

Abendanon, E.C., 1915, 1917, 1918. *Geologische en geographische doorkruisingen van Middle-Celebes*, 1909–1910, pt. I (1915), p. 1–451; pt. II (1915), p. 452–951; pt. III (1917), p. 592–1382; pt. IV (1918), p. 1383–1901. Leiden E.J. Brill

Ahlburg, J., 1913. *Versuch einer geologischen darstellung der insel Celebes.* Neue Folge Band 12, heft I. Jena, Gustav Fischer, 172 p., 11 pl.

Apandi, T., 1974. *Reconnaissance geologic map of Kotamobagu area, North Sulawesi.* Geol. Survey Indonesia, scale 1:250,000

Audley-Charles, M.G., D.J. Carter, and J.S. Milsom, 1972. Tectonic development of Eastern Indonesia in relation to Gondwanaland dispersal. *Nat. Phys. Sci.*, 239 (90), 35–39

Bemmelen, R.W. van, 1949. *The Geology of Indonesia*, vol. IA. Government Printing Office, The Hague, 732 p.

Brouwer, H.A., 1930. The major tectonic features of Celebes. *Proc. Kon. Akad. v. Weten.*, Amsterdam, 338–343

———, 1934. Geologische onderzoekingen op het eiland Celebes. *Ver. Geol. Mijnb. Gen. Ned. & Kol., Series*, 10, 39–171

———, 1947. *Geological exploration in the island of Celebes*, Geological summary and petrology. Amsterdam, 1–64

Carey, S.W., 1975. Tectonic evolution of Southeast Asia. *Proc. 4th Ann. Conv. Indon. Petrol. Assoc.*, 17–48

Djuri and Sudjatmiko, 1975. *Geologic map of Majene Quadrangle, South Sulawesi.* Geol. Survey Indonesia

Effendi, A.C., 1974. *Reconnaissance geologic map of Minahasa area, North Sulawesi.* Geol. Survey Indonesia, scale 1:250,000

Gribi, E.A. Jr., 1973. Tectonics and oil prospects of the Moluccas, Eastern Indonesia. Proc. Reg. Conf. on the Geol. of SE Asia, 1972. *Bull. Geol. Soc. Malaysia*, vol. 6, 11–16

Hamilton, W.H., 1970. *Tectonic map of Indonesia, a progress report.* US Geol. Survey, Denver, Colo, 29 p.

———, 1973. Tectonics of the Indonesian Region. Proc. Reg. Conf. on the Geol. of SE Asia, 1972. *Bull. Geol. Soc. Malaysia*, vol. 6, 3–10

———, 1979. Tectonics of Indonesia region. *Geol. Survey Prof. Paper.* US Govern. Print. Office, Washington, 345 p.

Kartaadipoetra, L.W. and T.W. Sudiro, 1982. A contribution to the geology of Southeast Sulawesi. *Geologi Indonesia* (Journ. Indon. Assoc.

Newfoundland Appalachians and the evidence for their transportation. A review and interim report. Geol. Assoc. Canad. Proc. (A Newfoundland decade) **24**, 9-25.

-----, 1973. Bay of Islands, map area, Newfoundland. Geol.Surv. Canad. Paper **72-34**, 7 pp.

-----, Turner, F.J. & Gilbert, C.M., 1954. Petrography. San Francisco : Freeman, 406 pp.

-----, and Malpas, J., 1972. Sheeted dikes and brecciated dike rocks within transported igneous complexes, Bay of Islands, western Newfounland. Canad. J. Earth Sci. **9**, 1216-1229.

Wilson, J.L., 1969. Microfaunas and sedimentary structures in 'deep water' lime mudstones. In : Friedman, G.M. (ed.) Depositional Environments in Carbonate Rocks. SEPM. Spec. Pub. 14, 4-19.

-----, 1975. Carbonate Facies in geologic history. New York : Springer-Verlag., 471 pp.

Wilson, C.J.N., 1980. The role of fluidization in the emplacement of pyroclastic flows : An experimental approach. J. Volcan. geoth. Res. Amsterdam **8**,231-249.

Winn, R.D.Jr and Dott, R.H.Jr., 1979. Deep-water fan-channel conglomerates of Late Cretaceous age, southern Chile. Sedimentology 26, 203-228.

Winterer, E.L. & Bosellini, A., 1981. Subsidence and sedimentation of Jurassic continental margin southern Alps, Italy. Am. Assoc. Petrol. Geol. Bull. **65**, 394-421.

Wiryosujono, S. & Hainim, J.A., 1975. Cenozoic sedimentation in Buton Island. Proc. Reg. Conf. Geol. Min. Res. SE. Asia, p. 109-119.

Woodcock, N.H., 1979. Sizes of submarine slides and their significance. J. Struct. Geol. 1, 137-142.

Woodcock, N.H. and Robertson, A.H.F., 1981. Wrench-related thrusting along Mesozoic-Cenozoic continental margins : Antalya Complex, SW Turkey. In: McClay, K.R. & Price, N.J. (eds.) Thrust and nappe Tectonics. Geol. Soc. London Spec. Publ. **9**, 359-362.

-----,-----, 1984. The structural variety in Tethyan ophiolite-terrains. In : Gass, I.G., Lippard, S.J. & Shelton, A.W. (eds.) Ophiolites and Oceanic Lithosphere. Geol. Soc. London Spec. Pub. 13, 321-332,

Worsley, T.R., 1979. Sea-level fluctuations and deep-sea sedimentation rates. Science **203**, 455-456.